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**EXTREME EVENTS
AND THE TRANSFORMATION OF LANDSCAPE**
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GUEST EDITORIAL

THE ROLE OF EXTREME EVENTS AND HUMAN ACTIVITY IN THE TRANSFORMATION OF LANDSCAPE: THE PHYSICAL GEOGRAPHY CONTEXT

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It is a great pleasure to be guest editor for this volume which honours 50 years of scientific activity by Professor Leszek Starkel. My pleasure as honorary editor arises from my affection for Poland, my great admiration for its geographers, and for their contributions, particularly over the last 3 decades since I first visited Poland in 1971. I have met Professor Starkel on numerous occasions, dating from our first meeting when the 20th International Geographical Congress met in London in 1964 – perhaps a little ironic that I chair the organizing committee for the 30th Congress also destined to meet in the UK, based in Glasgow in 2004, and that one of the themes will derive from an ICSU project on Past Hydrological Events and Global Change, in which Professor Starkel is a major research scientist. This present collection of papers, although focused on human activity and extreme events, ranges across the spectrum of several areas of physical geography, so that it is pertinent to provide some physical geography context in this editorial.

The last 50 years, the period during which Professor Starkel has been undertaking research, has been a particularly formative time for physical geography and the environmental sciences, not only for the ways in which the approach to environmental research has changed, but also for the way in which the public attitude to science and the environment has altered, with environmental awareness increasing. I have a very high regard for the manifest and brilliant contributions that Professor Starkel has made, but it would be presumptuous of me to try to

Table 1. The Development of physical geography

Changing Nature of Physical Geography, 2000		A century of British Geography, 2003	Some specific developments
Foundations 1850-1950	Uniformitarianism Evolution Scientific method Exploration and survey Conservation Cycle of erosion, zonal soils, climatic climax	Foundations up to 1900	Geology and Geological Survey foundation; Evolution; Glacial theory; Physiography; Cycle of erosion
		Evolution up to 1960	Denudation chronology Evolutionary theme in climatology and plant and animal biogeography Glacial and coastal geomorphology Experience of air photographs during WWII led to dynamic climatology Soil surveys
Developments 1950-1980	Positivism and post-positivism Quantification Data collection, Remote sensing	Revolution challenging earlier orthodoxy after 1960	Foundation of BGRG (British Geomorphological Research Group) in 1960 Quantitative and statistical methods
		Theory and modelling	Mathematical models Non linear modelling
Current Approaches	Environmental Systems	Processes	<i>Short term change</i> <i>Impact of individual events</i>
	Environmental Processes	Environmental Change	Increasingly precise dating Correlation of spatial sequences <i>Identification of short term sequences or individual events in the palaeoenvironmental record</i>
	Environmental Change	Human impact	Range of impacts identified <i>Timing of impacts, reaction and relation times</i>
	Human Activity	Applied and applicable research	<i>Hazards and extreme events</i> What is natural and basis for restoring nature
Trends for the Millenium	Global physical geography	Restructuring Global, GIS	Biogeochemical cycles Global change scenarios Upscaling and downscaling
	Cultural physical geography	New focii, cultural differences affecting environment, environmental management	Aesthetics of environment Holistic approaches, Human ecology Multidisciplinary approaches e.g. Geoarchaeology, Ecohydrology,
More integrated physical geography		Development of core – physical geography or physiography Relation to human geography Internal organization – geomorphology, climatology, hydrology, biogeography and ecological process, Quaternary environmental change Holistic and global problems, <i>global change</i> Relations with other disciplines – including impact assessment, risk analysis, environmental health, disaster management, urban regeneration, sustainable development, natural resource management	

Developed from Gregory, 2000, 2003a. Research on extreme events and human activity was an integral part of the specific developments listed and is indicated by italics.

summarise all those contributions and the impacts that they have achieved. Leszek has influenced the growth and evolution of the discipline, not only by his research contributions and publications, but also through the way that he has coordinated and directed research groups, has taken national and international initiatives, and has become very well known and prominent in the international academic arena. Professor Starkel has exerted a national and international presence in physical geography and has contributed to the development of the discipline in a very positive, creative and formative way.

I am attempting here to provide a brief, personal perspective on trends in physical geography over the last 50 years and to show how some of Professor Starkel's contributions fit into this context. I first attempted a review of the development of physical geography in 1985 (Gregory 1985); fifteen years later (Gregory 2000), what began as a revision, developed into a completely new book because so much had happened in the last fifteen years of the twentieth century as explained in a paper highlighting the changing nature of physical geography (Gregory 2001). In the British context, a review of the evolution of physical geography and environmental science, so inextricably bound up with developments elsewhere (Gregory 2003a) inevitably demonstrates a similar pattern of development. It is useful to reflect on the progress of physical geography in order to provide a context for studies of human impact and of short term events, allowing us to see why emphasis has been placed upon these themes in research, and so to contemplate where present trends should be leading us.

A summary of the development of physical geography (Table 1) shows the sequence of chronological phases recognized (Gregory 1985, 2000, 2001, 2003a), with the first two columns indicating trends for physical geography in general and the third showing those identified for Britain. This portrays the pattern of development of physical geography in terms of a sequence of approaches built upon foundations established in the period prior to the middle of the twentieth century (Gregory 2004). These sequential approaches succeeded the period in the 1960s when physical geography was revolutionized by the advent of quantitative and statistical methods, and then by mathematical models and by the change in focus from an idiographic to a nomothetic approach (Johnston 1979; Gregory 1985). During this period of nearly 50 years, the separate approaches identified in Table 1 all prospered and, initially, developed independently from others. In order to remedy lack of knowledge of landscape forming processes empirical field and modelling approaches were developed in research, reinforced by publication of important books. Investigations of environmental change were also undertaken independently, having grown from investigations of the Pleistocene and the Quaternary and from glacial geology. The theme of human activity attracted increasing interest in the late 1960s as it became realised that human impacts had affected environmental processes and landscapes so substantially that research was required to establish the ways in which such human impacts had taken effect. This led readily to a fourth theme complementing processes, environmental change and human impacts, that of applications

of physical geography which, through the investigation of hazards and of mapping of environments in terms of its classification and evaluation, enabled a further clear approach to physical geography to develop.

Although it is impossible to do justice to the breadth and substance of Professor Starkel's contributions, it is striking that he has made singularly influential contributions in the development of all four areas. In relation to environmental processes, his contributions on the role of extreme (catastrophic) meteorological events in the contemporary evolution of slopes (Starkel 1976) have continued to the present day, including investigations in Himalaya (Starkel and Basu 2000). In environmental change, investigations of specific areas such as the Wisloka valley (Starkel 1981), the seminal, collaborative work on the Vistula valley (Starkel 1982, 1987c, 1990, 1991a, 1995, 1996), and the general investigation of Quaternary evolution of landscapes in Europe (Starkel 1983) have all been extremely significant and internationally recognized research contributions. In addition there have been investigations of human impact in which an early paper was very influential in establishing the context (Starkel 1966), followed by specific analyses of the significance of human impact in Europe (Starkel 1987a,b). Applications of geomorphology and physical geography have emerged from many of these works, such as research on palaeohydrology (Starkel 1991b), but were particularly associated initially with the development of geomorphological mapping in which Poland made substantial and fundamental contributions: detailed maps have always been a prime feature of the research undertaken by Professor Starkel. In addition to contributions in these separate fields it is notable how, as his research has progressed, it has also evolved so that the boundaries between the fields have become less pronounced. This is very clear in the contributions that Professor Starkel made in establishing international research in palaeohydrology, first in the temperate zone as one of the two leaders of IGCP Project 158 (Starkel, Gregory and Thornes 1991) and then globally (Starkel 1989) as the first president (1991-1995) of the Global Continental Palaeohydrology Commission (GLOCOPH) of INQUA (1991-2003). This palaeohydrological research produced important progress in understanding the diversity of river valley evolution in Europe (Starkel 1995) and in improving the approach to long-distance correlation of fluvial events (Starkel 1991c). A further theme which has emerged from the productive fusion of several of Professor Starkel's research interests has been on short term hydrological changes: demonstrated in Europe (e.g. Starkel 2001) and more widely, where it is suggested that short term changes show how reconstruction from the past can be of significant value in interpreting aspects of future global change (Starkel 2003). In these, as in many other contributions that he has made, Professor Starkel has revealed the breadth of his knowledge, his ability to see the relationships and correlations from one place to another and from one time to another, and he has also always been aware of the need to involve all relevant disciplines – a truly multidisciplinary, imaginative and innovative scientist.

In the course of development of the separate approaches (Table 1), physical geography has necessarily fragmented as reductionism increased, but at least two important over-riding trends became increasingly apparent by the turn of the twentieth century. First, that new and additional approaches have become increasingly evident, which may be termed global and cultural physical geography (Gregory 2000). A more global approach was encouraged as a reaction to reductionist investigations of small detailed areas and problems, by awareness of the need to demonstrate the global context, by availability of higher resolution remote sensing data and more advanced geographical information systems, and by increasing realization of the potential of global and climate change. A cultural approach to physical geography (Gregory 2000, 2001), although in existence for more than four decades, was catalyzed by the growing need to consider public opinion and cultural attitudes in environmental management, and was fostered by progress in studies of human impact and their relationship with applications of physical geography. Second, all the separate approaches have continued to evolve but have each reached a point at which they need to interrelate with other approaches. Indications of the need for such interrelationships, and of multidisciplinary research, are tentatively suggested in the themes shown in the final column of Table 1. Some of these interrelationships have, in addition, produced a number of more separate trends, listed at the bottom of the final column in Table 1. One of these trends is that, as there has been decreasing clarity between the traditional branches of physical geography of geomorphology, climatology, and biogeography, there has been a resurgence of a holistic approach and a recognition of the need to focus upon the total physical environment and upon links with the human environment (Gregory, Gurnell and Petts 2002) presaging a restructuring of the discipline in which a renaissance of a more integrated physical geography provides the most likely future direction. A further general development has been the realization that we need to return to a focus upon particular areas. After the revolution of the 1960s it was natural and inevitable that physical geography should progress towards more general models in order to establish fundamental generalizations but, on the basis of the progress achieved, it is inevitable that particular places should now resume their position on the physical geography agenda (Phillips 2001); this is vital when applications of physical geography are related to restoring nature and design of the physical environment (Gregory 2003b).

It can be seen (Table 1) that the study of extreme events has arisen from several of the themes identified and is a research topic of much contemporary interest. The papers in this volume illustrate many of the aspects of research on extreme events and their relation to human activity, all with considerable significance for future understanding of environment. Analysis of the changing incidence of precipitation events is initially required as shown by Niedźwiedz (chapter 1) and Prokop and Walanus (2); leading to investigation of the significance of recent events for physical landscape, including mass movements (3), alluvia-

tion in valleys (4), valley floor evolution in a proglacial valley (5) and gully evolution (6). It is then desirable to see the way in which extreme events need to be considered in longer term evolution as illustrated for southern Germany (7) and Estonian landscapes (8). We have to employ specific investigations to progress to the general perspective and to applications which arise: these are demonstrated here by extreme events in the context of late Quaternary environmental change (9), by the extremeness of extreme events against earlier ideas of magnitude and frequency concepts (10), leading naturally to the relevance to environmental disasters (11) and finally to applications of the results, particularly in geomorphology (12). This sequence, from investigations of contemporary data records (1-2), to their relevance to recent environmental change, especially involving human activity (3-8), and to the implications that arise (9-12), reflects the way in which investigation of extreme events can aid our understanding of landscape. Professor Starkel's contributions have featured at all of these levels. I would like to join the contributors to this special volume who, I am sure, all wish their papers to celebrate the numerous contributions that Leszek has made to our understanding of landscape systems and to acknowledge their appreciation of his leadership and manifest support.

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EXTREME PRECIPITATION EVENTS ON THE NORTHERN SIDE OF THE TATRA MOUNTAINS

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PART I REGIONAL INVESTIGATIONS

ABSTRACT: This study reviews the interdecadal and decadal variability of extreme precipitation on the northern slope of the Tatra Mountains (southern Tatra), using monthly records of duration from 1 to 20 days. Daily data from 1947-1992 are used for the Tatra Mountains (the mountainous area) and for the Tatra Plateau (the non-mountainous area). This is compared with the nearest station in Poland. The highest daily maximum was 24.1 mm (1992) recorded on 20 June 1972 during a storm with cyclonic circulation. This is equal to the mean extreme value of precipitation from observed during 100-year years. In July 1984 the highest 5-day total reached 442 mm, and during the 15 days between 24 to 28 July 1984 the total amount of rainfall reached 769 mm. In the 1950s, when the precipitation series was the shortest, all extreme events are distinctly greater than in the period span of the extended 20-year period, although a strong reduction of the frequency of the data from the northern part 2001. However, an earlier period with extreme precipitation was observed during the years 1934-1977. A maximum 10-day total amount of extreme precipitation was 1216. However, there is no sign for the loss of the interdecadal variability of extreme values compared the values typical for interdecadal variation in the 1950s, which could mean that it is not so important a phenomenon as it is thought.

KEY WORDS: extreme events, extreme precipitation events, Tatra Mountains, Poland

INTRODUCTION

The Tatra Mountains are the highest part of the Western Carpathians and are influenced strongly by the heavy precipitation in southern Poland (Cielieba 1992). Sometimes, especially during the summer season, extreme rainfall

PART I
REGIONAL INVESTIGATIONS

EXTREME PRECIPITATION EVENTS ON THE NORTHERN SIDE OF THE TATRA MOUNTAINS

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ABSTRACT: This article reviews the occurrence and variability of extreme precipitation on the northern slope of the Tatra Mountains (southern Poland), using rainfall amounts of duration from 1 to 30 days. Daily data from 76 years (1927-2002) are used for the Hala Gąsienicowa meteorological station at 1520 m a.s.l. This is the place with the heaviest rainfall in Poland. The highest daily precipitation total (300 mm) was recorded on 30 June 1973 during a northern cyclonic situation. For longer durations extreme values of precipitation were observed during different years. In July 1934 the highest 3-day total reached 422 mm, and during the 11 days between 16 to 26 July 2001 the total amount of rainfall reached 500 mm. In the last 7 years the precipitation totals and the number of extreme events are distinctly greater than in the previous part of the analysed 76 years period, although a strong influence on the results may be the data from the extreme year 2001. However, an earlier period saw extreme precipitation concentrated during the years 1958-1978. A transition to a rather wetter phase of climate has been noted since 1995. However, there is no sign for any of the elements studied of any departure that has exceeded the values typical for fluctuations of climate in the 20th century, and which could therefore be taken as indication a permanent change in the climate.

KEY WORDS: climatic change, extreme precipitation events, Tatra Mountains, Poland.

INTRODUCTION

The Tatra Mountains are the highest part of the Western Carpathians and are influenced strongly by the heavy precipitation in southern Poland (Niedźwiedź 1992). Sometimes, especially during the summer season, extreme rainfall

amounts cause catastrophic floods in the large area of the upper Vistula basin. During the 20th century more than 41 significant flood events occurred in this region caused by the strong precipitation in Carpathian Mountains (Cebulak 1998, Cebulak and Niedźwiedź 2000). The best known events took place in July 1934, 1970 (Niedźwiedź 1972) and 1997 (Niedźwiedź 1999). According to the orographic effect the northern slope of the Tatra range is especially affected by extremely high precipitation with daily totals achieving 300 mm (Cebulak 1983, 1992a), which is the highest value ever measured in the whole of Poland (Paszyński and Niedźwiedź 1999).

Since 1995 the frequency of local heavy rains connected with thunderstorms has probably increased in southern Poland (Cebulak and Niedźwiedź 1997). After prolonged rains in July 1997 the new record level of summer precipitation occurred in summer 2001. All these extreme precipitation events play an important role on morphogenetic processes (Kłapa 1980), and in intensifying the denudational system in the mountains (Kotarba et al. 1987) causing large amounts of erosion even in the form of debris-flows (Kotarba 1998). The intensity of these phenomena seems to be as large as those during the Little Ice Age (Starkel 1996, 1999). There are some suggestions that the increase in the number of extreme events in the mountains is connected with a general acceleration of energy and mass circulation (Starkel 1999) caused by the increasing greenhouse effect (Obrębska-Starkel and Starkel 1991, Bednarz et al. 1994).

The main aim of this paper is investigation of the variability of selected extreme precipitation events in the Tatra Mountains during the last 76 years. I hope that a preliminary assessment is possible, about whether precipitation increased and whether extreme events have become more frequent during recent years or not.

MATERIALS AND METHODS

The most representative place for the northern side of the Tatra Mountains with a relatively long precipitation data series is selected for analysis to meet the aims of this paper. Daily data from 76 years (1927-2002) are used for the Hala Gąsienicowa meteorological station (49° 15'E, 20° 00'E) located at 1520 m a.s.l., strictly near the upper tree line. This is the place with the heaviest rainfall in Poland. The station is operated by the Institute of Meteorology and Water Management, as well as by the Institute of Geography and Spatial Organization, and data quality is very good. Measurements of daily precipitation total started in December 1926. But during the war there are breaks in the data: August-December 1939, January 1940, July-August 1944, and the whole year 1945.

Statistical analysis was possible for 27565 days. For each year were calculated the extreme values of precipitation totals for durations from 1 to 30 consecutive days. Also standard monthly and annual totals, as well as the winter (December-

February), spring (March-May), summer (June-August) and autumn (September-November) precipitation sums were analysed. Frequency, empirical probability and standard time series analyses of data were performed.

EXTREME PRECIPITATION

Hala Gąsienicowa has an average annual precipitation of 1690 mm which varied from 1038 in 1946 to 2626 mm in 2001 (Table 1). Above 42 percent of the annual total is observed during the summer, what is typical for this part of Europe. Spring precipitation (23 percent) is slightly above that of autumn (22 percent), and the winter precipitation is the least (13 percent). Variability of the presented element is greatest during the autumn season. Maximum monthly precipitation occurs in July (250 mm) with a relatively large coefficient of variability (55 percent). Extreme monthly totals varied from 38 mm in July 1928 to the exceptional value of 743 mm in July 2001. The highest monthly value for the whole of the Carpathians was 812 mm measured at the Lysa Hora peak in the Silesian Beskid Mountains in the territory of the Czech Republic in July 1997 (Niedźwiedź 1999). Amounts exceeding 500 mm were noted on four occasions: in July 1934 (684 mm), 1980 (622 mm), 1997 (560 mm) and 1960 (518 mm).

Table 1. Average and extreme precipitation totals (in mm) in Hala Gąsienicowa (1927-2002).

Element	Winter (DJF)	Spring (MAM)	Summer (JJA)	Autumn (SON)	Annual	Daily Maximum
Average	220	387	714	368	1689	90,0
Coeff. of variation (%)	32	27	30	32	17	51
Maximum	469	675	1439	672	2626	300,0
Year	1948	1940	2001	1931	2001	1973
Minimum	92	94	381	160	1038	34,7
Year	1997	1946	1990	1986	1946	1935

The highest daily precipitation total (300 mm) was recorded on 30 June 1973 during a northern cyclonic situation. This is the highest 24-hour amount of rainfall ever observed in the Carpathian Mountains (Cebulak 1983) as well as in the whole of Poland. The highest daily value for Central Europe was 345,1 mm on 30 July 1897 (Paszyński and Niedźwiedź 1999) in the Isera Mountains (Sudetes) at the Nova Louka (Neuwiese) station in the Czech Republic, and 313 mm in the Zinnwald (Erzgebirge) near the Czech-German border on 12 August 2002. Such large and prolonged rains are influenced by the orography, when the humid air masses are flowing perpendicularly to the mountain chain. All investigations indicate that such events are connected with the northern, north-eastern, and north-western cyclonic situations or cyclonic troughs (Cebulak 1992b, Lapin and Niedźwiedź 1984, Niedźwiedź 1972, 1999).

For longer durations, extreme values of precipitation were observed during different years. In July 1934 the highest 3-day total reached 422 mm. During 11 days between 16 and 26 July 2001 the total amount of rainfall reached 500 mm. Another wet period was observed on 26 June – 18 July 1934 with 685 mm during 23 days. The extreme value of 700 mm was exceeded on 26 days and 779 mm on 30 consecutive days (Table 2).

Table 2. Extreme precipitation totals (in mm) in Hala Gąsienicowa (1927-2002) for duration 1-30 days.

Duration days	Precipitation mm	Period	Duration days	Precipitation mm	Period
1	300,0	30 June 1973	16	592,0	25 July – 9 August 1991
2	392,5	16-17 July 1934	17	614,0	11-27 July 2001
3	422,4	16-18 July 1934	18	634,0	1-18 July 1934
4	438,7	15-18 July 1934	19	647,4	30 June – 18 July 1934
5	462,3	14-18 July 1934	20	656,9	29 June – 18 July 1934
6	465,3	13-18 July 1934	21	662,3	28 June – 18 July 1934
7	467,2	12-18 July 1934	22	679,0	27 June – 18 July 1934
8	473,2	11-18 July 1934	23	684,7	26 June – 18 July 1934
9	473,2	10-18 July 1934	24	685,1	22 July – 14 August 1980
10	482,7	14-23 July 1934	25	696,5	21 July – 14 August 1980
11	499,5	16-26 July 2001	26	700,1	21 July – 15 August 1980
12	561,6	16-27 July 2001	27	740,1	1-27 July 2001
13	570,2	15-27 July 2001	28	744,9	1-28 July 2001
14	575,8	14-27 July 2001	29	772,3	20 June – 18 July 1934
15	575,8	13-27 July 2001	30	779,4	19 June – 18 July 1934

Apart from prolonged extreme events sometimes the greatest erosion consequences have followed violent heavy rains caused by local thunderstorms. In the Tatra Mountains the maximum rainfall of 60 minutes duration exceeded 40-50 mm with a probability 1 percent (return period 100 years), and 30 mm with a probability 10 percent (once in 10 years). However, the maximum values for rainfall duration exceeded 60-80 mm with 1 percent frequency, and 40 mm with 10 percent frequency (Cebulak et al. 1986, Niedźwiedz 1986c, 1992).

Daily precipitation exceeding 200 mm was recorded three times during the 76 years (Table 3), but more than 100 mm was observed 25 times. Long term variability of daily maximum precipitation (Figure 1) indicates a small increasing trend (1,5 mm for 10 years) from 84 mm in 1927 to 95 mm in 2002, but the highest values occurred during the 1958-1978 period.

The most evident is variability of the number of days with precipitation above selected thresholds (Figure 2). For example the number of days with precipitation >10,0 mm changed from 49 in 1927 to 53 in 2002 with the small increasing trend of 1 day for a 20 years. The maximum number of such days was observed in 1948 (74), with a secondary maximum in 2001 (67 days).

Table 3. The number of days (n) with precipitation above particular thresholds in Hala Gąsienicowa, in relation (percent) to 27565 days of observation (1927-2002).

Threshold mm	Number of days	Percent	Threshold mm	Number of days	Percent
0,1	16161	58,6	80,0	61	0,2
1,0	12481	45,3	90,0	38	0,14
5,0	6821	24,7	100,0	25	0,09
10,0	3840	13,9	110,0	17	0,06
20,0	1543	5,6	120,0	12	0,04
30,0	754	2,7	130,0	10	0,04
40,0	404	1,5	140,0	7	0,03
50,0	239	0,9	150,0	4	0,015
60,0	150	0,5	200,0	3	0,011
70,0	96	0,3	300,0	1	0,004

For days with larger precipitation amounts the highest number was noticed in 2001. In this particular year the number of days with precipitation above 50 mm exceeded 13, and above 30 mm exceeded 22 days.

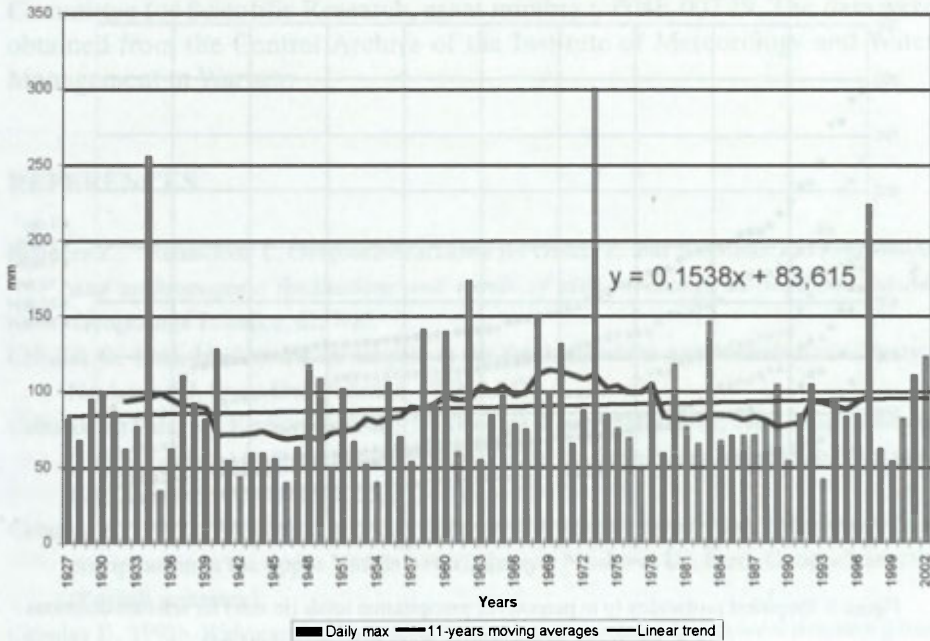


Figure 1. Long-term variability of daily maximum of precipitation in Hala Gąsienicowa.

The empirical probability (Figure 3) of long lasting precipitation enabled the evaluation of risk of occurrence of such precipitation. For example with the

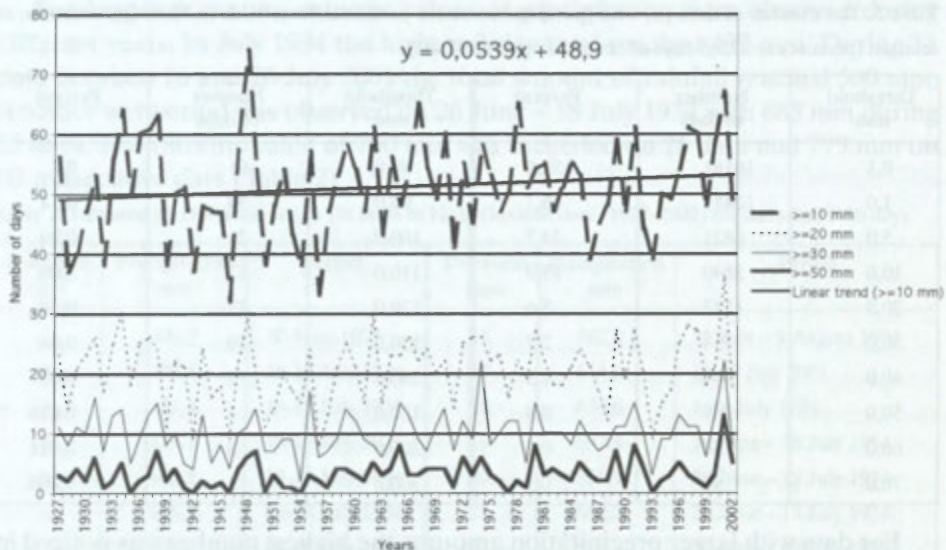


Figure 2. Variability of the number of days with precipitation above selected thresholds in Hala Gąsienicowa.

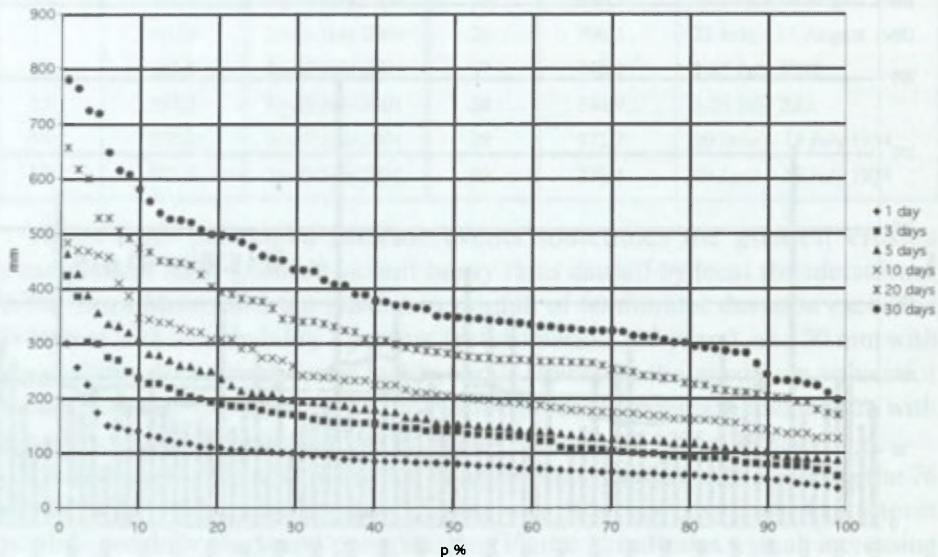


Figure 3. Empirical probability (p in percent) of precipitation totals (in mm) for selected durations (1, 3, 5, 10, 20 and 30 consecutive days) in Hala Gąsienicowa.

return period of 10 years ($p=10$ percent) daily precipitation exceeded 130 mm, 3 days precipitation exceeded 230 mm, 5 days – 300 mm, and 30 days total of precipitation could be higher than 600 mm.

CONCLUDING REMARKS

The problem of variability of large precipitation totals during the last 76 years on the northern slope of the Tatra Mountains was studied. There exists a large dispersion of extreme events of different duration. Extreme values exceeded 300 mm in 24 hours and near 800 mm on 30 days.

In the last 7 years precipitation totals and the number of extreme events is distinctly greater than in the previous part of the 76 year period analysed. It may be that the data from the extreme year 2001 has a strong influence on the results. But it is generally evident that the previous period with extreme precipitation was concentrated during the years 1958-1978. A transition to a rather wetter phase of climate has been noted since 1995. However, there is no sign for any of the elements studied of any departure that has exceeded the values typical for fluctuations of climate in the 20th century, and which could therefore be taken as indicating a permanent change in the climate.

ACKNOWLEDGEMENTS

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TRENDS AND PERIODICITY IN THE LONGEST INSTRUMENTAL RAINFALL SERIES FOR THE AREA OF MOST EXTREME RAINFALL IN THE WORLD, NORTHEAST INDIA

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ABSTRACT: The longest instrumental rainfall series from any region in the North Asian subcontinent and 7 surrounding tropical stations in Northeast India. Analysis attempts to reveal and estimate rainfall trends in the study area, with an emphasis on the period over North Asian during the last 130 years. The Fourier analysis of fluctuations in rainfall series shows that the generally equal of 1977 year is the dominant one with a signal, with the same phase, has been found for all rainfall stations of the North Asian subcontinent and in all investigated series.

KEY WORDS: extreme rainfall, trend, periodicity, northeast India

INTRODUCTION

Studies of the annual/seasonal (or other exact) amount of variability in rainfall over India have a long history. Blanford (1860) first prepared annual rainfall series for British India, while Walker (1916) first analyzed the longest continuous series for the period 1841-1905, and reported that no trend could be observed. Later many authors analyzed rainfall series for periods of different length using data from different rain gauge networks. Parthasarathy and Mookhy (1978) found that no trend was to be noted for the all-India annual maximum rainfall series. By analyzing 1971-1988 monsoon rainfall data, Subbaranyaya and Naidu (1982)

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ABSTRACT: The longest instrumental rainfall series have been investigated for the North Assam subdivision and 4 meteorological stations in Northeast India. Analysis of trends to annual and seasonal rainfall show these to be very stable, with no change in the rainfall over North Assam during the last 150 years. The Fourier analysis of fluctuations in rainfall series shows that the periodic signal of $T=3.5$ year is the strongest one. Such a signal, with the same phase, has been found for all rainfall stations of the North Assam subdivision and in all investigated seasons.

KEY WORDS: monsoon rainfall, trend, periodicity, northeast India.

INTRODUCTION

Studies of the annual/summer (south-west) monsoon variability on rainfall over India have a long history. Blanford (1886) first prepared annual rainfall series for British India, while Walker (1910) first analyzed the summer monsoon series for the period 1841-1908 and reported that no trend could be observed. Later many authors analyzed rainfall series for periods of different length using data from different raingauge networks. Parthasarathy and Mooley (1978) found that no trend was to be noted in the all-India annual monsoon rainfall series. By analyzing 1871-1988 monsoon rainfall data, Subbaramayya and Naidu (1992)

reported downward trends for rainfall during the late 19th century and the 1960s.

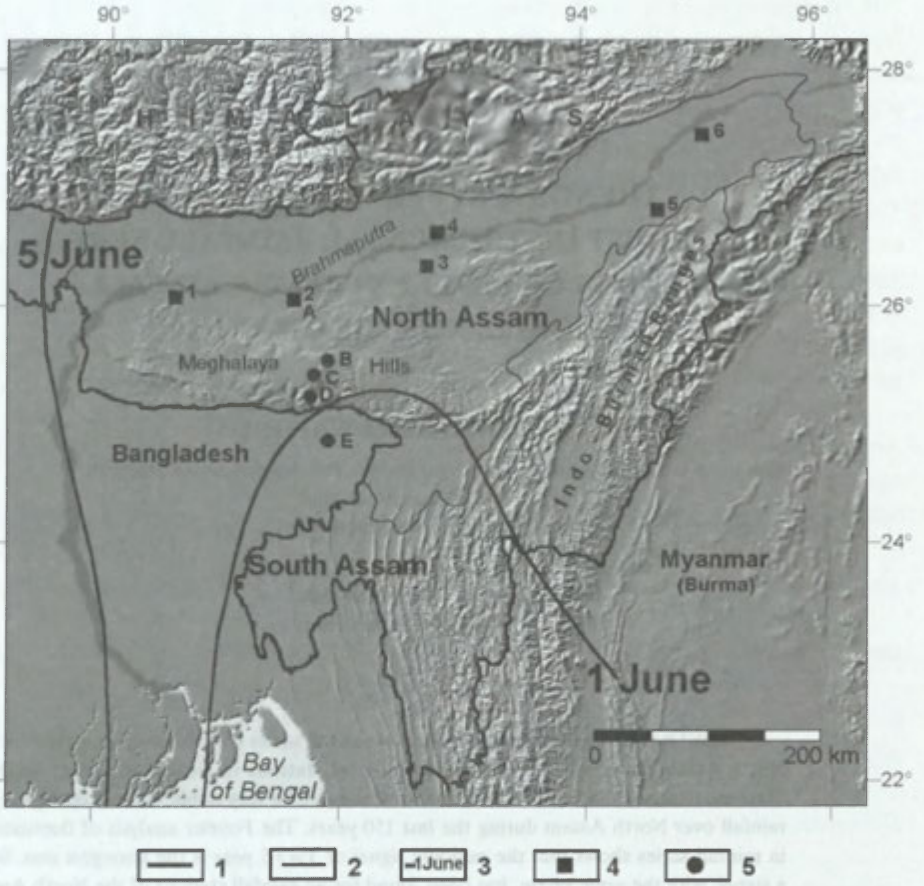


Figure 1. Location of Northeast India region with North Assam subdivision and Meghalaya Hills.

- 1—national border, 2—subdivision boundary, 3—location of rain gauge stations used by Parthasarathy et al. (1995) for the North Assam subdivision average rainfall calculation: 1 - Goalpara, 2 - Gauhati, 3 - Nowgong, 4 - Tezpur, 5 - Sibsagar, 6 - Dibrugar, 4 - stations used for detailed trend and periodicity analysis along profile: A—Gauhati, B—Shillong,
- C—Mawphlang, D—Cherrapunji, E—Sylhet (station in Bangladesh added only to show orographic effect);
- 4—normal dates of onset of the south-west monsoon (after Rao 1981).

Most studies therefore indicate that monsoon rainfall has been mainly random in nature over a long period of time, particularly on the all-India scale (Moley and Parthasarathy 1984). At the subdivisional and station scales the presence of some periods (especially in the range 2-3 years) have been found to be significant (Parthasarathy 1984). Unfortunately the source of these oscillations is not known and cannot be linked to the El Nino-Southern oscillation (ENSO)

in the tropics (Bhalme and Jadhav 1984) or to oscillation of stratospheric winds (Bhalme et al. 1987).

The northeast area of India (north of 21°N and east of 88°E, Figure 1) can be treated as a relatively separate macroregion (Winstanley 1973, Parthasarathy et al. 1987). Gregory (1989) found that linear trends indicating a reduction in rainfall come close to achieving significance over this area. Sontakee and Singh (1996) reported that summer monsoon rainfall is negatively but weakly correlated with that of Peninsular India and Rupa Kumar et al. (1992) found significant downward trends for monsoon rainfall over the period 1871-1984.

Detailed studies show differences within a region between two North and South Assam subdivisions, when it comes to rainfall trends and periodicity. Parthasarathy and Dhar (1974) noted an upward trend over North Assam and a negative one over South Assam for the period 1901-1960. Bhaskar et al. (1998) found significant downward trends for the North Assam subdivision for the periods 1871-1888, 1918-1962, and a significant upward trend for the period 1888-1918. The same authors noted a downward trend for South Assam for the years 1951-1994, and an upward trend for the years 1928-1951.

Studies at individual stations of North Assam have concentrated mainly on the abnormally high rainfall at Cherrapunji (Blanford 1889, Starkel 1972, O'Hare 1997, Singh and Syiemlieh 2001, Starkel et al. 2002). Cherrapunji has the world record for high rainfalls over durations of between 31 days and two years since 1860-61 (WMO 1986). Most authors agree that orography is the main cause of the enormous rainfall at this station.

DATA AND METHODOLOGY

The mean monthly rainfall data for the North Assam subdivision (1871-1999) have been taken from the published dataset of Parthasarathy et al. (1995). Additionally, long instrumental monthly mean rainfall data for the Cherrapunji (1872-2000), Shillong (1869-2000), and Gauhati (1848-2000) observatories were collected from the National Data Center, India Meteorological Department (IMD), Pune, and from the Assam State Electricity Board (ASEB) in the case of Mawphlang station (1899-1987). The data for North Assam were used as a background for a detailed analysis of annual and seasonal monsoon trends and periodicity along the profile Gauhati-Cherrapunji. The position and altitude of rain-gauge stations were verified using GPS and an altimeter during fieldwork in 2000.

Four climatic seasons are distinguished over northeast India (Rao 1981): winter (January-February), the pre-monsoon (March-May), the south-west monsoon (June-September) and the post-monsoon (November-December). The trends and periodicity of rainfall time series have been investigated for a whole year, and for the two seasons of the southwest monsoon, when 70 percent of pre-

precipitation falls, and of the pre-monsoon, when cyclonic storms form over the Bay of Bengal. The thunderstorms connected with the latter produce large amounts of rain (mango rains) in April and May, and are typical for northeast India (Pant and Rupa Kumar 1997).

The rainfall series have been statistically tested using Student's *t*. In the course of exploratory periodicity analysis of rainfall time series, a Fourier analysis was performed with the emphasis on a maximal periodic signal of $T=3.5$ year.

NORTH ASSAM RAINFALL PATTERN AND ITS SIGNIFICANCE

The investigated North Assam subdivision covers the two states of Assam and Meghalaya and has a total area of 56,339 km². The average annual rainfall calculated from 6 stations for the period 1871-1999 reached 2,226 mm (Figure 1). However, subdivisions are administrative demarcations and do not represent internally cohesive areas. This can be seen clearly in the case of North Assam. This zone is characterised by an extreme contrast between the southern part (Meghalaya) with an annual rainfall of 11,000 mm at Cherrapunji station, and the Brahmaputra Valley (Assam) in which only 1600 mm annual rainfall is noted in Gauhati (Figure 2).

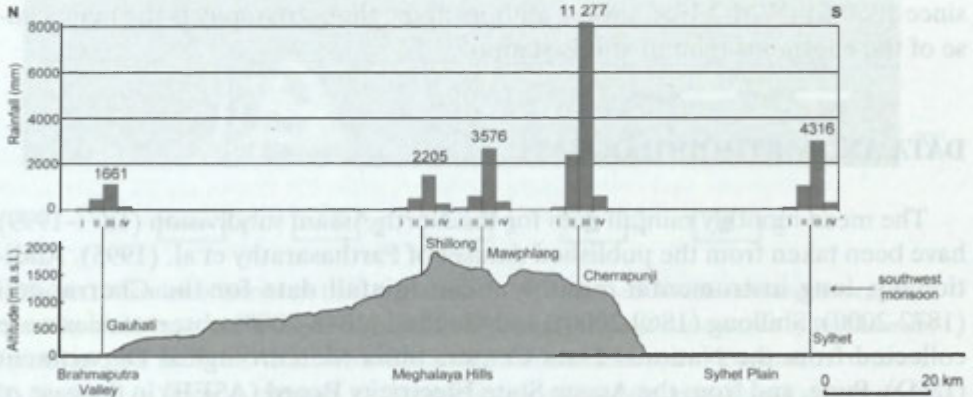


Figure 2. Cross section of the Meghalaya Hills with seasonal rainfall distribution for uniform period (1901-2000).

I—winter, II—pre-monsoon, III—south-west monsoon, IV—post-monsoon.

The Meghalaya Hills, a relatively small region located between the Brahmaputra Valley in the north and the Bangladesh floodplains in the south, plays an especially important role. They form the first orographic barrier for the humid southwest monsoon winds on their way from the Bay of Bengal. The Meghalaya Hills account for about 20-25 percent of rainfall input during spring and in June

even though they represent only 2 percent by area of the Ganges, Brahmaputra and Meghna basins (Hofer 1997). The rainfall over the southern slopes of the Meghalaya Hills is very important, perhaps even decisive, in the flood processes noted in Bangladesh (Hofer, Messerli 1997). Dhar and Nandargi (1998) also rate the north-east among the flood-prone areas in which a majority of floods in India occur.

THE TREND TO RAINFALL DATA

Visual inspection of rainfall time series suggests that no evident trend is present in the data (Figure 3). The numerical, statistical, approach to the question of the trend is as follows. In line with Ockham's razor, the simplest approach is used. The data series is divided in half and the simplest statistical test (Student's t) is used to answer a question as to whether on average, the first and second halves of the series differ. The results are given in Table 1.

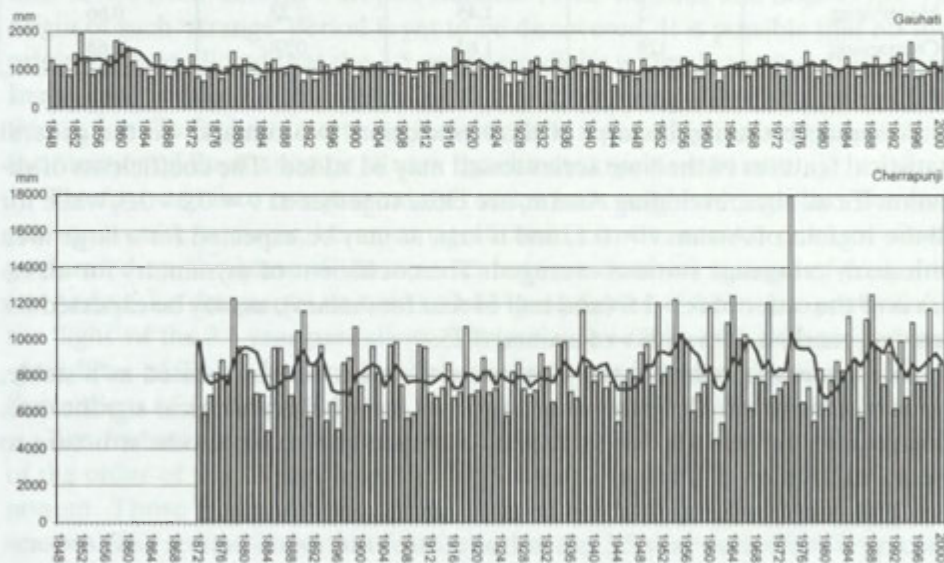


Figure 3. Plots of the south-west monsoon rainfall series at the two selected stations Gauhati (Brahmaputra Valley) and Cherrapunji (Meghalaya Hills). Continuous curve shows 5 years moving average.

The values of t are generally small (under the null hypothesis of no trend, the expected value of t is 0, and its standard deviation 1). The sign to t -values may have some geographical meaning, since the positive trend is obtained for stations in close proximity Shillong, Mawphlang and Cherrapunji, while the Gauhati and North Assam subdivisions have a negative trend. It must be mentioned, however, that the negative t values are by no means significant statistically. The highest

obtained positive value $t = 2.23$, if taken as a number alone, gives a relatively significant p-value of 0.030. However, taking into account the fact that 2.23 is the extreme of 12 values, the significance of the value drops significantly. The average $t_{av} = 0.658$ ($n=15$), gives a significant p-value=0.01. However, the assumption of independence of 15 values of t , necessary for such a conclusion, is evidently broken by the fact that both the North Assam and Gauhati values are of negative sign, while all others are positive. The final conclusion is that there is no trend to rainfall data in the region.

Table 1. Student's t-values for difference between later and earlier half of time series.

Raingauge station	Series length (years)	Year	Pre-monsoon	South-west monsoon
North Assam (6 stations average)	129	-0.44	-0.51	-0.51
Gauhati	152	-0.28	-0.63	-0.46
Shillong	134	2.01	1.25	1.26
Mawphlang	75	1.45	2.23	0.66
Cherrapunji	129	1.63	0.58	1.63

To enhance the significance of that conclusion, two remarks on the general statistical features of the time series tested may be added. The coefficients of variation for all sites, excluding Assam, are close together at $v = 0.2 - 0.3$, while for all the regions of Assam $v = 0.1$, and is less, as may be expected for a large area with many raingauge stations averaged. The coefficient of asymmetry for all series is of the order of 0.5 -1.0 (and half of that for Assam), as may be expected for positive, random data with v of around 0.25.

As a general conclusion the given rainfall series may be treated as a stable, random time series. Only the average values are of climatological significance. However, in the next section an attempt is made to find a periodic structure to the rainfall data.

PERIODICITY: THE PERIODIC SIGNAL OF $T = 3.5$ YEAR

In the course of the exploratory data analysis of the rainfall time series, a Fourier analysis has been performed. For the statistical significance assessment it is important to mention that the plot on Figure 4 is the first obtained in data exploration. It is clear that the periodic component with a period $T=3.5$ year is of maximal amplitude. The horizontal scale is that of the period of the periodic component. Maximal amplitude is obtained for $T=3.5$ year. The second one is about $T=11$ year, although the evidence for this is too weak to connect rainfall with the

sun's activity. Similarly, the 3.5 year period seems not to be statistically significant, anyway that is of maximal amplitude.

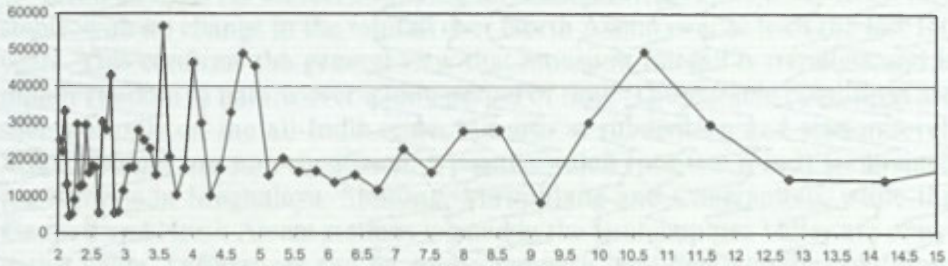


Figure 4. Fourier analysis result of rainfall time series for southwest monsoon season (June-September) at Cherrapunji station. On the horizontal axis the period T is plotted (the scale is not linear for T because it is linear for frequency, i.e., the reciprocal of T).

The 3.5 year period is known to be present in many rainfall and river runoff time series from Central Europe (Walanus 1990, Walanus and Soja 1995). The origin of such 'strange' period is yet to be discovered. It is possible that no such origin exists at all and that the 3.5 year periodicity is simply random fluctuation. In any case, it would seem interesting to search for the spatial distribution of this fluctuation. To find more time series with such a period is the only way to track the mentioned fluctuation or to seek the climatological cause of it.

The first step is to compare rainfall series from neighbouring sites. The comparison would, however, be more significant from the natural point of view if performed on the basis of the time scale (time domain) instead of frequency (or period; i.e. frequency domain). On Figure 5 twelve series are presented in the 'light' of the 3.5 year periodicity. The series are filtered by a Gaussian band-pass filter of $T = 3.5$ year; relatively narrow but wide enough to retain visible fluctuations of amplitude of the 3.5 year component, over tens of years.

Two features of plots are examined. The first is the existence of fragments of the order of the 20 year length of higher amplitude of 3.5 year periodic component. Those fragments are, more or less, synchronous over sites and over seasons. The second important feature is that the 3.5 year periodicity is synchronous itself.

The sites are close together, so such parallelism of (filtered) series may be expected. Nevertheless, it proves that the periodicity discussed is not a random fluctuation of one site's rainfall, but is of, at least regional significance.

What is interesting is that both kinds of synchronisms appear between the non-overlapping 'pre-monsoon' and 'southwest monsoon' seasons. It means that rainy phases, within a 3.5 year period, are visible not only in one 'season' but also in all. It seems to agree with the conclusion of a regional (global?) rainfall correlation with a 3.5 year periodicity.

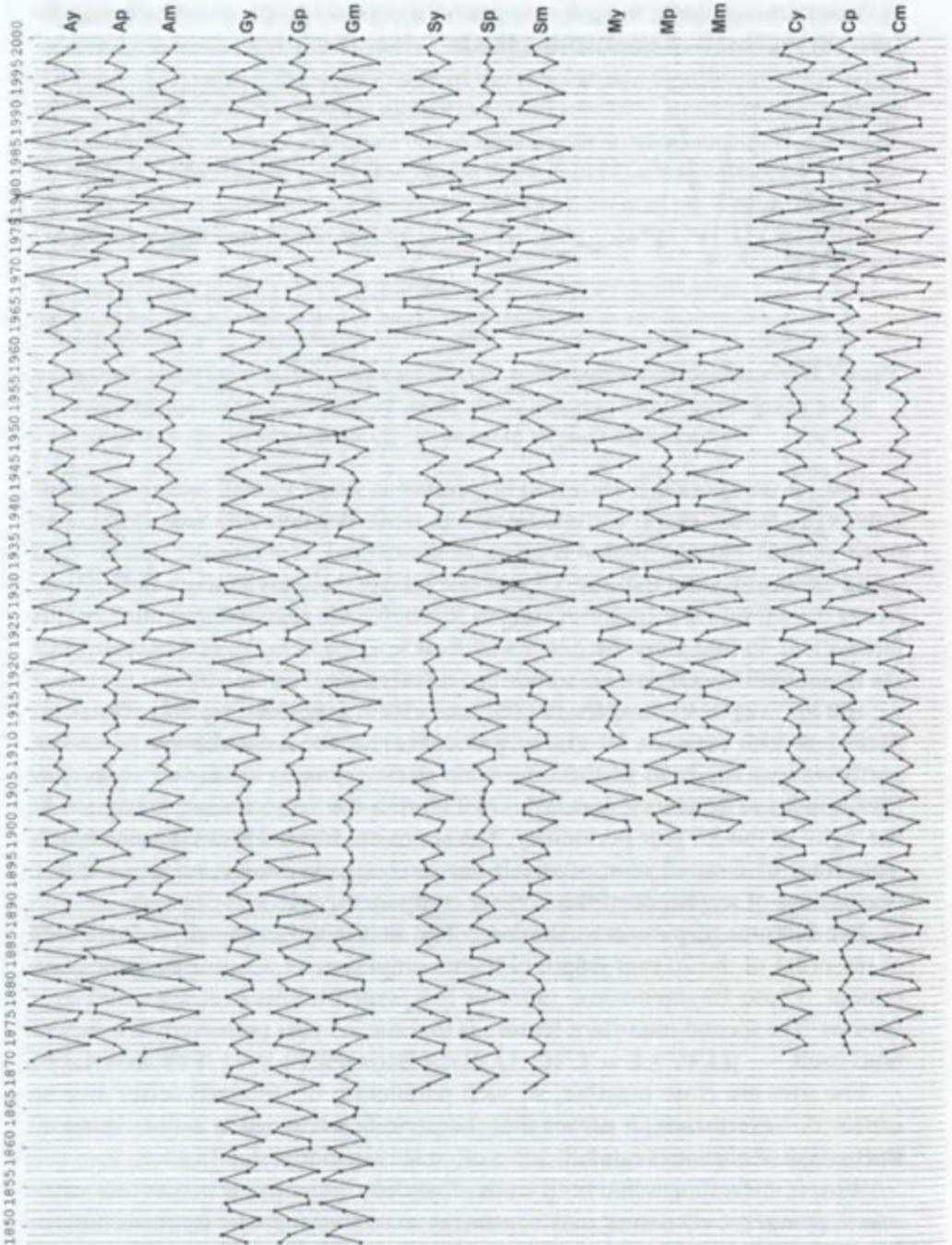


Figure 5. Filtered rainfall time series, i.e. presented in the 'light' of 3.5 year periodicity.

Stations: A—North Assam subdivision, G—Gauhati, S—Shillong, M—Mawphlang,

C—Cherrapunji; seasons: y—year, p—pre-monsoon, m—southwest monsoon.

The filter is a relatively narrow Gaussian band-pass filter of $T=3.5$ year.

CONCLUSIONS

Trend analysis for annual and seasonal rainfall series shows these to be very stable, with no change in the rainfall over North Assam over at least the last 150 years. This confirms the general view that monsoon rainfall is trendless and is mainly random in nature over a long period of time. These stable conditions are seen not only on the all-India scale, but also at subdivision and station level. While t-values are non-significant, a positive value (positive trend) is obtained for stations in Meghalaya: Shillong, Mawphlang and Cherrapunji, while the Gauhati and North Assam stations located in the Brahmaputra Valley are of negative value. Differences can be connected with modified climatic conditions over the Brahmaputra Valley, as located in the 'rain shadow' between two belts of high precipitation: the Meghalaya Hills and Indo-Burman Ranges in the south and the Himalayas in the north.

The analysis of fluctuations in rainfall series shows that the periodic signal of $T=3.5$ yr is the strongest one. Signals with the same phase have been found at all rainfall stations and in all investigated seasons over the North Assam subdivision. A similar signal is also known to be present in other parts of the world (e.g., central Europe) in rainfall and river runoff series. We can assume that, while the source of the signal is not known, the periodic component of $T = 3.5$ year is independent of local or regional climatic conditions over monsoon areas like India.

These conclusions do not change the fact that the monsoon is highly variable in intensity from year to year, and that extreme average precipitation values with a contrasting rainfall pattern exist in the North Assam. These have an important, especially hydrological, significance over a much larger area. The documented rainfall conditions stable in time can give a good basis for future research, e.g., on the influence of land use changes on soil erosion, the water balance or the river regime on different scales.

ACKNOWLEDGMENTS

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INTRODUCTION

With rapid urbanization and a streamlining of growth of tourism in Durgam and Nizampur (Figure 1) the landscape is undergoing an unprecedented rate of population and development. Consequently, extensive areas are being lost with the gradual elimination of vegetation. The scientific and practical aspects of land coupled with vulnerable geological structure and high rainfall have led to the establishment of vicious cycles of soil erosion and landslides. Durgam and

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CAUSES AND CONSEQUENCES OF LANDSLIDES IN THE DARJILING-SIKKIM HIMALAYAS, INDIA

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ABSTRACT: Landslides are common disaster phenomena in many countries causing great economic losses. The Darjiling-Sikkim Himalayas in India is known to be the most severely affected. Even a glance at landslide statistics gives some idea of the enormity of damage done and the ever present threat to lives and property. It has been observed that even 50 mm of rainfall in an hour would cause landslips. Unauthorized structures in the unsafe zones, absence of an adequate drainage system and unplanned growth of settlements have accelerated the process of ecological imbalance. In recognition of the acuteness of problems related to landslides, this paper summarizes our knowledge of slope stability so as to provide information on the origin of slope movements and the methods of their investigation, prevention and control. A series of case studies in the Darjiling-Sikkim Himalayas have been undertaken to provide a better understanding of this acute natural disaster problem.

KEY WORDS: disaster phenomena, inadequate drainage, landslide, slope stability, unsafe zone, India.

INTRODUCTION

With rapid urbanization and a phenomenal growth of tourism, the Darjiling and Sikkim Himalayas (Figure 1) have been experiencing an unprecedented rise in population since independence. Consequently, pressure on land is increasing with the gradual elimination of virgin forests. Unscientific and unplanned usages of land coupled with vulnerable geological structure and high rainfall have led to the establishment of vicious cycle of soil erosion and landslides. During or after

every monsoon, landslides create havoc in the Darjiling-Sikkim Himalayas devastating human lives and properties. As a result the Darjiling-Sikkim Himalayas, one of the most densely populated tourist centers in such monsoon environments, is on the verge of an environmental catastrophe so that just one concentrated shower of 50 mm per hour could initiate landslides endangering innumerable local people and their properties.

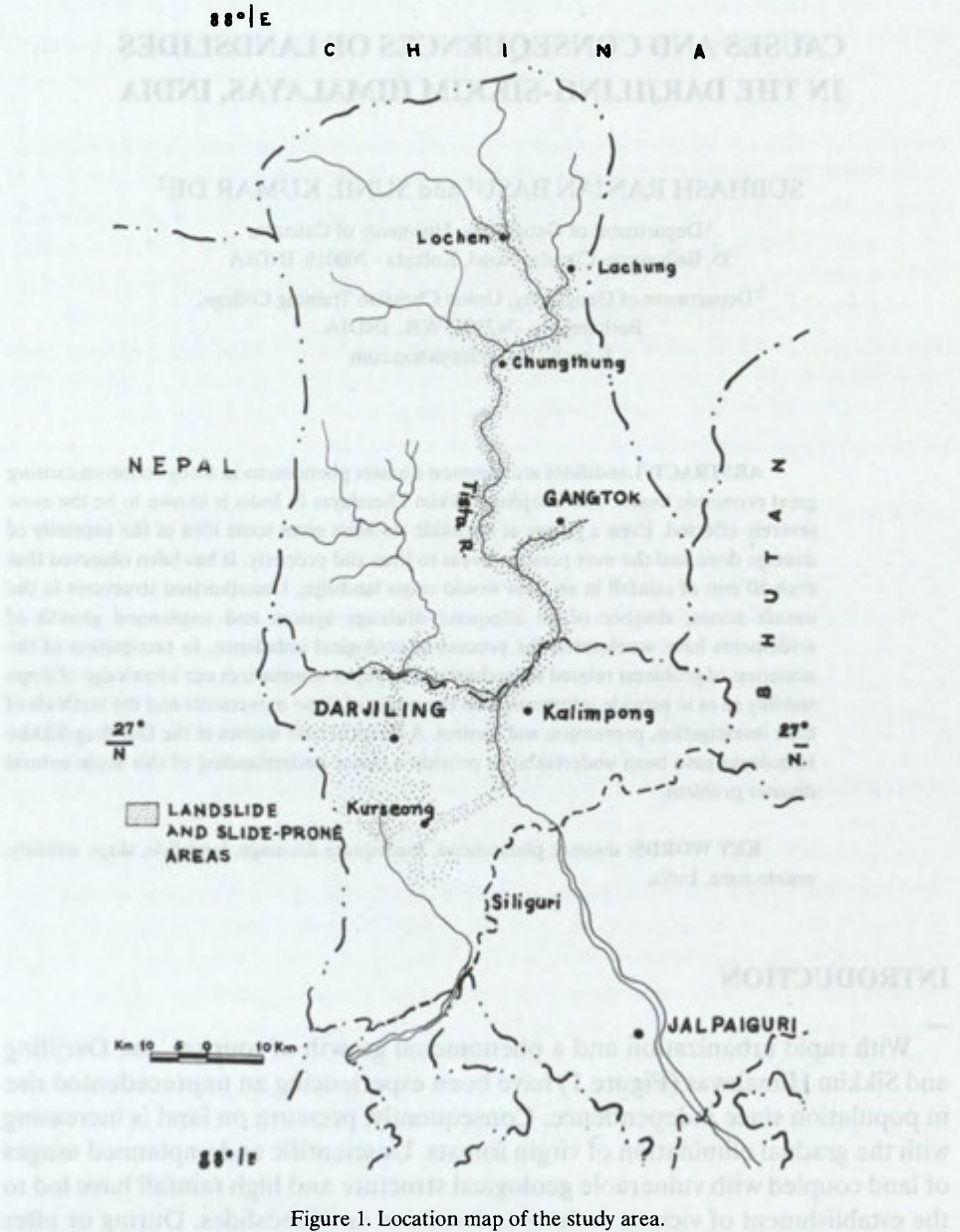


Figure 1. Location map of the study area.

The methodology adopted to identify the causes of this problem is rationalistic involving determination of slope instability, analysis of slope materials, detailed tabling of the composition and orientation of geological structure and the examination of geomorphological processes together with the nature and extent of human interference. Data have mostly been obtained through intensive field surveys and other published sources.

PREVIOUS WORK

In spite of the fact that vast amounts of geo-technical information have been collected as a result of demand-oriented studies carried out by researchers (Ghosh 1950, Nautiyal 1951,1966, Dutta et al. 1966, Roy and Sen Sharma 1967, Basu 1969-1970, Verma 1972, Paul 1973, Sinha 1975, Chatterjee 1983, Sengupta 1995, Ghosh 1998) of the Geological Survey of India, no attempt had so far been made to synthesize the data to obtain a regional picture of the landslide problem in Darjiling-Sikkim Himalayas. The present paper attempts to fulfill this objective by collating results of studies carried out by the authors and their associates and also by incorporating relevant observations of other workers, which are duly acknowledged at the appropriate places in the text.

MAJOR LANDSLIDE EVENTS

DARJILING HIMALAYAS

Available records show a disastrous landslide occurred on the 24th September, 1899, in and around Darjiling town due to unprecedented rainfall of 1065.50mm, killing 72 persons and causing enormous loss to land and property (Griesbach 1899-1900). Most of these fifteen slides were confined to the soil-material overlying the gneissic rocks. The instability of the hillsides gradually increased due to progressive absorption of moisture from excessive rainfall and the cutting of hill slopes both for natural and artificial needs together with defective drainage.

The second major event of landslips in Tindharia, Darjiling, Kalimpong and Kurseong towns took place on the 15th January, 1934, due to Bihar-Nepal earthquake, which was responsible for widespread destruction though not of equal magnitude to that experienced in 1899. On this occasion, the top layers of the sub-soil on the crest of the Darjiling ridge and its outlying spurs, mostly on the western side of the town, developed fissures damaging buildings.

Between the 11th and 13th June, 1950, the hill slopes in and around Darjiling, Kurseong, and Kalimpong towns were badly affected by a series of landslides after a heavy spell of rain of 834.10mm causing widespread damages to roads, railways, houses and public works. 127 people were killed and several hundreds

were rendered homeless. The Siliguri-Kalimpong railway line was closed forever, as the hillsides in that region were considered unsafe for railways.

All these events pale into insignificance in comparison with the appalling calamity, which overtook Darjiling-Himalayas late in 1968. Due to incessant and heavy rain of 1121.40mm between 3rd and 5th October, 1968 there were numerous landslides accompanied by unprecedented floods in the Tista and other rivers. Hill-Cart Road and NH 31 were completely washed away. Several bridges at strategic points (The Rongpo bridge on the Sikkim border, a one span concrete bridge at Tista bazaar (market), the Railway bridge at Sevok and several others) were either washed away or severely damaged. The death-toll, officially estimated was 677 while unofficial reports placed the figure much higher. The landslides at Giddapahar near Kurseong, damaged over 175m of road and railway track and demolished many *bustee* (slum) hamlets. All the roads in Kalimpong town were badly battered and many houses collapsed due to land subsidence. According to the Indian Tea Association, 10-15 percent of the total tea area in the Darjiling Himalaya was destroyed together with the loss of 100 lives and widespread damage to factory-buildings and other installations.

During 3rd and 4th September 1980, due to heavy rainfall of 299.1mm Rim-bik, Lodhama, Bijanbari, Darjiling town, Sukhiapokri, Manebanjan, Sonada, Tindharia, Happy Valley, and Ambootia (Kurseong) were affected by severe landslides, killing 215 people and destroying properties worth about 100 million rupees. 462.5 mm of heavy rainfall during 15 and 16th September 1991 created numerous landslides in and around Darjiling town, Ging, Tukvor, Bennockburn, Bloomfield tea gardens and Pagla *jhora* and Chunabhata areas killing 2 people and severing the railway connection between hills and plains. It took almost 5 months to restore the railway connection between Darjiling and Siliguri.

In 1993 due to the occurrence of heavy and concentrated rainfall of 211.3 mm during 11th - 13th July, innumerable landslides devastated Mangpoo, Takdah, Pesoke, Rangtong, Tindharia, Pankhabari, Mahanadi, Gayabaari, Ambootia, and Darjiling towns. 15 people were killed in Mangpoo alone, and damage to properties was quite severe.

The years of 1995, 1998, 2001, 2002 and 2003 saw the latest cases of landslips in Darjiling town and along the Hill-Cart road from Kurseong to Darjiling. Due to high rainfall (300 - 600 mm) on the 5th - 7th July 1998, the discharge of all the streams increased so greatly that they cut the toe of the Hill-Cart road, causing numerous landslides, toe-erosion and subsidence.

SIKKIM-HIMALAYAS

In Sikkim Himalayas the landslides of Gangtok town, the Capital of Sikkim, are the only ones studied in detail, as the town has recently been severely affected by landslides due to the ever-increasing pressure of big buildings catering for the hotel trade associated with the enormous rise in the tourist population.

Landslides in Gangtok Town

The first active landslide cum subsidence was reported from the Chandmari area of Gangtok town. It was quite active during 1975-76 and thereafter remained relatively stable till June, 1984 when it was reactivated. Another slide at Tathangchen first occurred in the 1975 monsoon affecting the hill slope below Tathangchen village adjoining Chandmari. During the 1984 monsoon, this old slide was reactivated together with additional areas further south of the old slide.

The next recorded landslides occurred in the Syari and Deorali areas of Gangtok Town on the 16th September, 1990 due to a heavy rainfall of 85 mm. There were nearly 25 deaths and considerable loss of properties especially of the army cantonment. The next landslide disaster was on the night of the 5th September, 1995 in the Deorali area of Gangtok town, killing 30 people and damaging lot of properties.

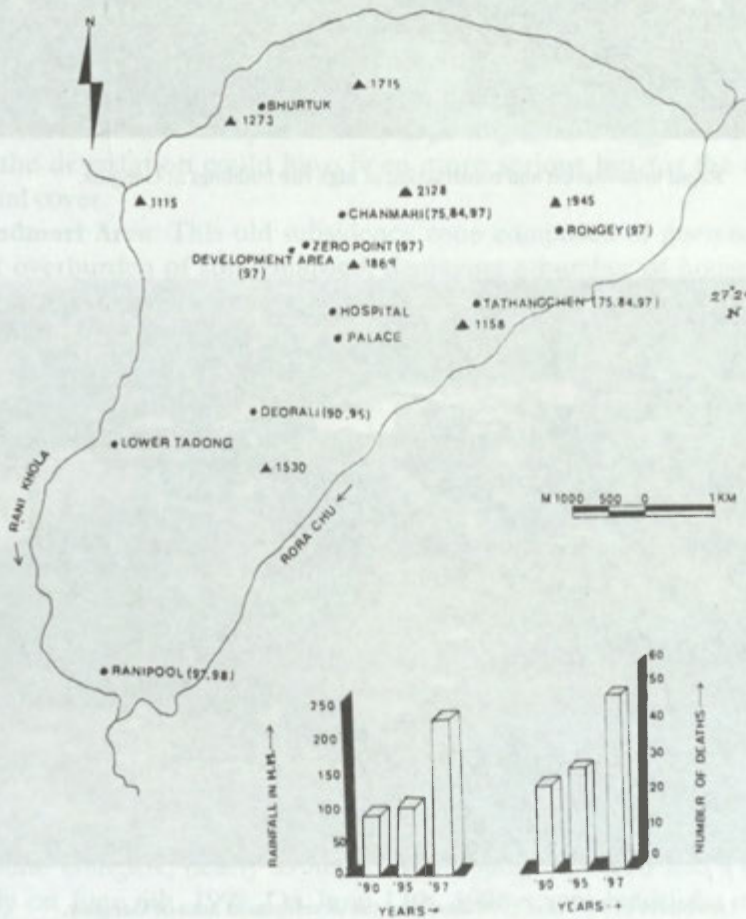


Figure 2. Location of landslides and No. of deaths in Gangtok during 1975-1998.



Rapid urbanization and construction of high rise buildings at Gangtok.



Landslides of 1997 and 1998 disrupted the development area of Gangtok.

Plate 1.

On the night of the 8th June, 1997, due to a heavy downpour of about 220 mm a disastrous series of landslides gripped the town of Gangtok and its suburbs, killing 50 people and devastating a large number of houses. The details of the localities severely affected (Figure 2) where:

- **Zero Point:** Area lying close to the Raj Bhavan and landslide destroyed the duty-huts and killed the constables on-duty. The dislodged materials reached up to the office complex of the Institute of Cottage Industries of NH 31A. The affected slope is composed of schist and gneissic rocks with a thick overburden of soil. The slope seemed to have undergone rotational failure.
- **Development Area:** Sudden subsidence of a portion of NH 31A, following heavy flow of rainwater on the road due to blocking of drains, caused collapse of multistoried houses and cottages leading to the highest number of casualties at a single location and also destroying a large number of parked vehicles. Under-scouring of the schistose and phyllitic structure by the accumulated water was recognized as the main cause of the ultimate collapse.
- **Tathangchen Area:** This area experienced the rotational failure of a slope with thick soil cover leading to some deaths, destruction of houses and livestock. The sudden burst of a septic tank upslope might have triggered this calamity and the devastation could have been more serious but for the fairly good vegetal cover.
- **Chandmari Area:** This old subsidence zone composed of gneissic rock with thick overburden of soil collapsed, damaging a number of houses, a school complex, a temple, a government vehicle workshop and a large number of parked vehicles.
- **Rongey Basti (Donkan Dara) and Chongey Tar Basti on Gangtok-Rongey-Bhusuk-Assam (GRBA) Road:** The road suffered about 20 minor and major landslips as well as formation washouts. At Rongey Basti, two houses collapsed and some livestock perished. Changey Tar Basti suffered extensively with one death and collapse of 14 houses and destruction to other properties. Only one *pucca* (brick built) house survived, withstanding the main impact of falling debris. The 45 villagers, who took shelter in this *pucca* house, were thus saved.

In the year 1998 the entire state of Sikkim was affected by the rainfall induced landslides which began in the month of May but continued for entire June and July with severe results leading to heavy loss of lives, properties, irrigation canals, mini-hydel power units, roads and bridges. A preliminary assessment of monetary losses carried out by the Government of Sikkim in the middle of July, 1998 indicated a total loss of approximately Rs. 1,370 million. The worst situation was in the East Sikkim with 40 percent of its area affected in some form or other.

Around Gangtok, nearly 65 houses were badly damaged and a further 118 partially on June 6th, 1998. On June 13th, 1998 severe landslides at Ranipool caused death to 3 persons and serious injuries to 4 others when a few multi-storied houses collapsed due their impact.

CAUSES OF LANDSLIDES

It has been observed that landslips generally follow a period of heavy rains, which gives the final pressure when other factors have already brought the slope nearly to the point of failure. Since in a slip, the material is sheared off the parent body and moves in a downward direction, it is evident that the force causing the slip is the weight of the material and gravity; and during the time of slipping the component force of gravity down the slope has exceeded the resistance to shearing and sliding in unconsolidated material and rocks forming the slope. This can result either due to undercutting of the slope by erosion or by human agencies, for example slopes are steepened by undercutting for laying roads, or developing flat terraces for building sites. Decrease in the shearing resistance of the material adjoining a slope is caused by progressive structural changes in the material. The most common of these is weathering and formation of clayey materials. Percolation of water also causes reduction in the shearing resistance of the material forming the slope. Thus, a soil or talus material which is ordinarily stable in a certain slope may cease to be so due to the decrease of the shearing resistance caused by undue addition of water during a heavy storm. In slopes formed of talus or rock debris, there is practically no cohesion between the different blocks, so that the stability of slope depends upon internal friction alone. The angle of internal friction in rock debris which is sometimes 30° to 40° , is known as the angle of repose. A slope will safely stand to a great height with an inclination slightly less than the angle of repose. In slopes formed of rocks, slips occur only along a certain planes of weakness like joints or bedding planes, a massive rock, free of such weak features may even stand vertically up to great heights (Dutta 1950).

In the Darjiling-Sikkim Himalayas landslides have been caused by one or all of the factors mentioned above. In this paper the devastating landslides of the 1998 along the Hill-Cart road from Lower Pagla jhora up to Darjiling Town have been studied in detail. In the case of Sikkim Himalayas the landslides of the Tathangchen and Chandmari areas have been detailed with diagrams.

LANDSLIDES AFFECTING HILL-CART ROAD DURING JULY, 1998.

The most common occurrence of landslides in the Darjiling Himalayas is found along the springs which invariably cut-across the roads connecting the hills with the plains. Geologically speaking the Daling rocks (phyllites, slates, schists, etc.) and the Damuda rocks (sandstone, shales interbedded with coal) are more susceptible to landslides. As such the present study has been restricted to the arterial routes to Darjiling, which attracts thousands of tourists every summer, and due to the landslide damage every monsoon, the economy and the ecology of the region have been seriously affected. Sometimes landslides lead to the total disruption of vehicular traffic between the hills and the plains with the consequent disastrous effect on the transport of goods and tourist operations.

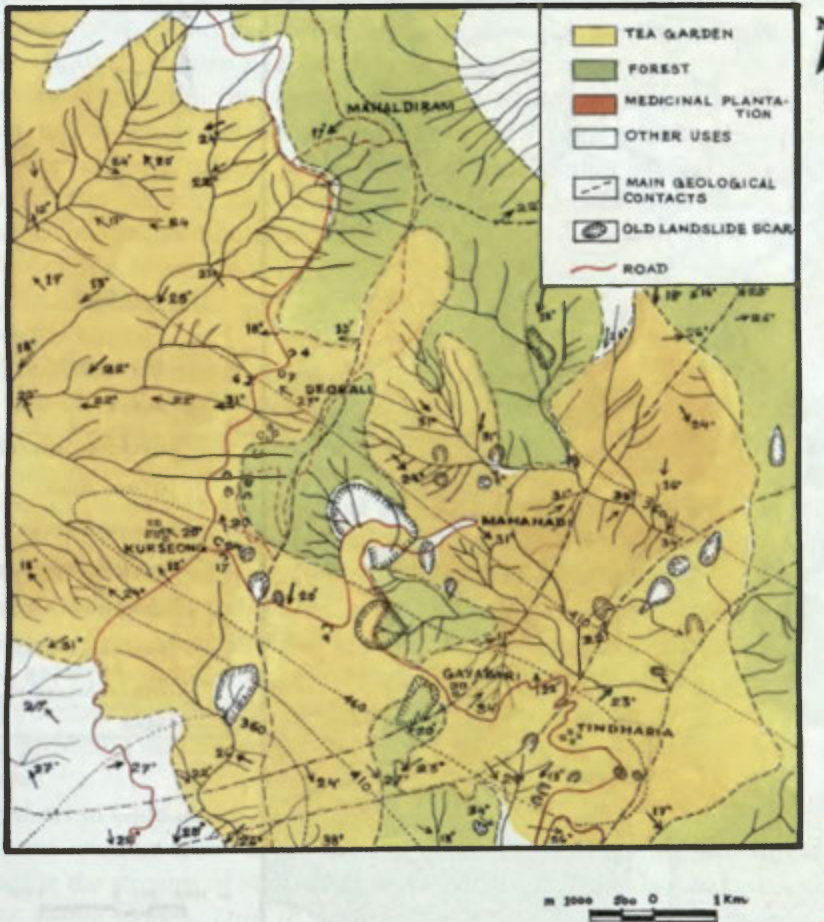


Figure 3. Map of geological units, landslides and landuse between Tindharia and Khairakhola.

To assess the nature of damage caused by devastating landslides during 5th to 8th July, 1998 in Darjiling Himalayas a geo-technical evaluation was attempted as follows.

The break in slope caused by the road cum railway bench provides the seat for deposition of the overflowing streams. During every monsoon the road bench is covered with slope wash in most places. These slope wash materials consist of large boulders to fine silt. Down slope the discharge from the streams causes toe erosion, below the culverts and causeways. Similar erosion occurs at the valley slope in sections where high discharges flow over the road and where adequate drains are not provided. The impact of large rolled boulders also causes damage to the road and the railway line (Plate 2). Even when major instability of the upper slope is not observed the slope wash material causes disruption of traffic along the said road.

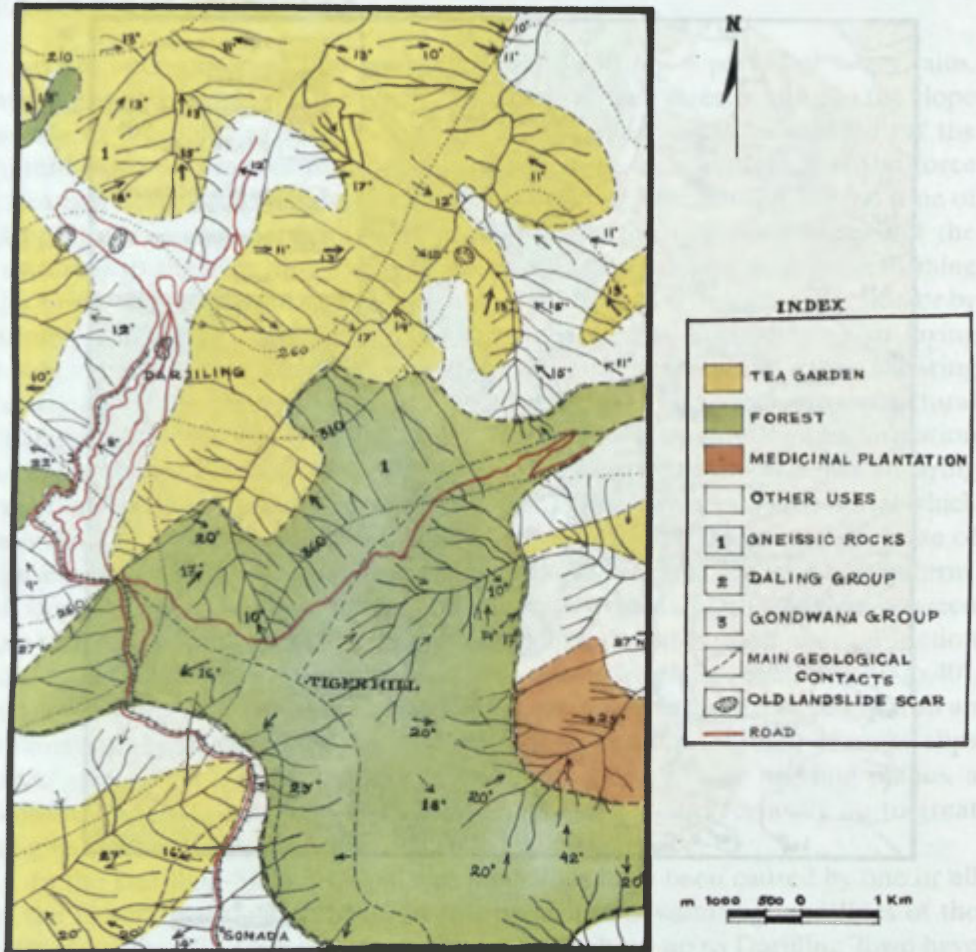


Figure 4. Map of geological units, landslides and landuse between Khaira-khola and Darjiling.

It has been observed that due to high rainfall during the 5th to 8th July, the discharge of all the streams increased greatly. The exceptionally high discharge, which could not be contained within its drainage channel caused major disasters as has been observed at the intersection of many streams with road sections.

Another major cause of landslides on the above mentioned areas is the down slope discharge of the streams which had been temporarily blocked at culverts until the rising water level of the *jhora* (spring) flowed along available depressions on the road bench to create a new path diverting the *jhora* away from the earlier established course. The new discharge channel cut the toe of the road causing landslide, toe erosion and subsidence (manifested by large arcuate cracks) on the road bench. The damage to the road bench in all cases has been proportional to the discharge of the *jhora*. In case where the *jhora* is oblique to

the road, the high discharge and consequent choking of the culvert, the abutments of the road had been damaged allowing the *jhora* to follow a new course. Such a feature can be observed south of Dilaram. It is therefore necessary to train the river upstream of the culverts so that the course of the *jhora* remains almost perpendicular to the road or else preferential erosion of one bank will take place. Observation indicated that where the discharge exceeded the carrying capacity of the *jhora*, the road bench also acted as a drainage channel, as mentioned above.

The major observed damage caused by landslide and subsidence that occurred along the Hill-Cart road (NH 55) may be grouped into the following four major sectors:

• **Chunabhati - Gayabari sector**

The Hill-Cart Road follows a steady gradient uphill from Chunabhati to Gayabari. The road bench, by and large, is free from any breach. The observed landslides near Tindharia are shallow, surficial debris slides occurring both on upslope and downslope of the road bench. Besides over-saturation of the soil, the slope instabilities may be related to human activities such as felling of trees, mining activity etc. The triggering effect was the high rainfall on the slope. With removal of the slide debris the problem in this sector may be temporarily overcome. However it has been observed that the contour drains along the road on the hill-slope slide are either choked at places or already damaged / cracked. Proper maintenance of the drainage system with reduction in slope in certain areas, providing flexible breast retaining walls, can ensure normal traffic along this road.

• **Gayabari to Giddapahr Sector**

The road section from about 37 km from Siliguri follows a water divide separating the streams of Siva Khola and Pagla *jhora*. There is rapid urbanization on the valley slopes. Due to high precipitation as well as surface run-off, infiltration on this valley slope has been intensified creating instability of both the railway track and the road bench. Subsidence cracks with sinking areas were observed indicating the initiation of a landslide. The Pagla *jhora* and its tributary from about 37.5km to 45.5 km have had devastating effects of the rail cum road. The high rainfall in its catchment broke a large number of existing culverts, overflowed on to the road, cut the valley slope to discharge its water downstream, changing its course (Plate 2). The villages like Mahanadi and the tea estates situated on the left bank have been totally cut-off both from Kurseong and Gayabari. The high rainfall in its catchment broke a large number of existing culverts, overflowed on to the road, cut the valley slope to discharge its water downstream changing its course. Data indicates that on 7th and 8th July, 1998 the rainfall at Dow hill was 151 mm and 436 mm, respectively compared to Kurseong which were 313mm and 32 mm, i.e., an increase of 242 mm on the eastern slope compared to the western slope. Preliminary morphological assessment indicates that in and around Kurseong the western



Highly denuded and subsiding Hill-Cart Road at lower Pagla *jhora*.



Subsiding Hill-Cart Road at lower Pagla *jhora* and damage to a railway engine at upper Pagla *jhora* by large rolled boulders.

Plate 2.

slope is open and gentler whereas the eastern slope is more restricted, confined and steep. Thus, the very high surface discharge on the eastern slope is concentrated through Pagla *jhora* and its tributaries. As a natural process valley widening, erosion of the toe of the bank and head-ward erosion mentioned above took place in the main *jhora* and its tributaries and consequently a large number of slides originated on the upslope due to increased instability. To accommodate this excess rainfall the road bench provides an oblique water channel following the gradient of road. In addition, as mentioned earlier, due to a break in slope the road bench acted as a zone of accumulation for the debris. Water and the slide debris on the road bench flowed down the gradient cutting the valley slope. This new water channel eroded the slope and reduced the stability of the road bench by toe erosion. Prior to heavy discharge this road was partially choked by rolled boulders and the slide-debris upstream of the culverts totally blocked the drainage, causing a temporary storage which with the increased discharge cut across the road bench to the valley below. At both upper and lower Pagla *jhora* on the existing road bench large wide-open arcuate cracks were visible. Such areas would remain as 'sinking zones', ultimately leading to landslides involving a portion of the road bench (as its head). Infiltration of surface run-off into these cracks has been noted at many places. The increase of pore-pressure within the soil on valley slopes due to infiltration was also responsible for slope failures. The south-eastern slope of the Gidda pahar at about 46 km from Siliguri shows a surficial slope failure on the hill slope and subsidence zone towards the valley. These are indicative of the instability of the area and the possible presence of a slip circle below the road bench.

• **Kurseong to Khaira *khola* Sector**

The road between Kurseong and Khaira *khola*, a distance of about 3 km is free from any road breach except at St. Alfansus School where the culvert has been washed away and near the Amardeep Hotel. Between this sector there are about 5 small streams which intersect the road bench. Except for Khaira *khola*, north of St. Mary's Seminary, no major disaster has been observed. Some surficial landslides have been noted on the slopes of streams within the Kurseong municipality area. All these slides triggered by heavy rainfall were due to super-saturation of the slope material accompanied by toe-erosion of the high gradient streams. However at Khaira *khola* there appears to have been two road breaches on either side of the culvert. Observed from the road two debris slides containing various sizes of boulders, cobbles and gravels in a sandy/ silty matrix occurred upstream of the culvert. Here again prior to the road breach it may be assumed that existing culvert was completely choked allowing the surface discharge from the two streams to accumulate behind the culvert. As in the case of Pagla *jhora* when storage height exceeded the height of the road bench, two drainage channels on either side of the culvert were created. Water also continued to flow over the road into the cracks created by subsidence towards the valley slope.

- **Khaira khola to Darjiling Sector**

North of Khaira *khola*, except for a major breach in and around Dilaram in this sector of NH 55, seven landslides/subsidence have been observed. In most of these cases it was observed that the flow channel and culverts were choked by debris. Observation indicates that the culverts were not constructed perpendicular to existing roads thereby allowing the surplus water to flow over the road bench and induce toe-erosion. In such cases the surplus water was forced to follow the road drains, until it could cut the road bench and make a new flow channel. The toe erosion mentioned above damaged the breast walls and toe-support. At two places near Pakhrin Village, culvert no. C312 where the crown of an old slide existed, the toe-erosion caused subsidence of the road bench. Damages to the road at culvert no. C340 causing a road breach was due to toe-erosion during heavy and uncontrolled discharge. In this place the width of the chute-drain down slope was insufficient. The observed subsidence in this area was perhaps aggravated by the construction of an unlined 'sump' pit near the road bench on the upslope side.

TATHANGCHEN SLIDE COMPLEX OF SIKKIM HIMALAYAS

Three prominent slope failures occurred near Gangtok below Tathangchen affecting the eastern hill slope of the north-south trending Mintokang-Palace-Tashiling ridge. The affected area is drained by a number of easterly flowing perennial jhoras, e.g., Palace *jhora*, Tathangchen *jhora* and some unnamed *jhoras*. The three slides near the Tathangchen village have been mapped in Figures 5 and 6.

- **Palace *jhora* slide**

This is an old slide which had initially been caused by toe-erosion near the confluence of the Palace *jhora* with Roro chu (river). Presently the crown of the slide has reached about 30 m short of the road level on the eastern side of the Palace. Subsidence of the road formation adjoining the Palace is the culmination of the head-ward extension of this slide.

- **Tathangchen Minor slide**

This slide occurred during the 1984 monsoon affecting a small depression area located about 500m north-east of the Palace *jhora* slide. The failed slope consists of regolithic soil, a product of weathering underlying sericite-chlorite schist, which is highly crumpled and disturbed with a dip of 50-55 degrees towards the east. The slide has been caused by the valley-ward movement of over-burden material together with highly weathered bedrock along the plane of foliation that forms the dip slope.

- **Tathangchen Major slide**

During the 1984 monsoon this old slide (1975) was reactivated and additional areas further south were severely affected due to lateral extension of the slide. On a rough estimate an area measuring 800-1000 m in length and 200-250 m in width below the village was badly damaged. The debris from the adjoining

Chandmari slide rolling down the Chanmari *jhora* caused toe erosion on the right bank which has resulted in the failure of the Tathangchen hill slope. This slide is considered to be a major slope stability problem endangering the entire habitation of the eastern slope of Gangtok town.

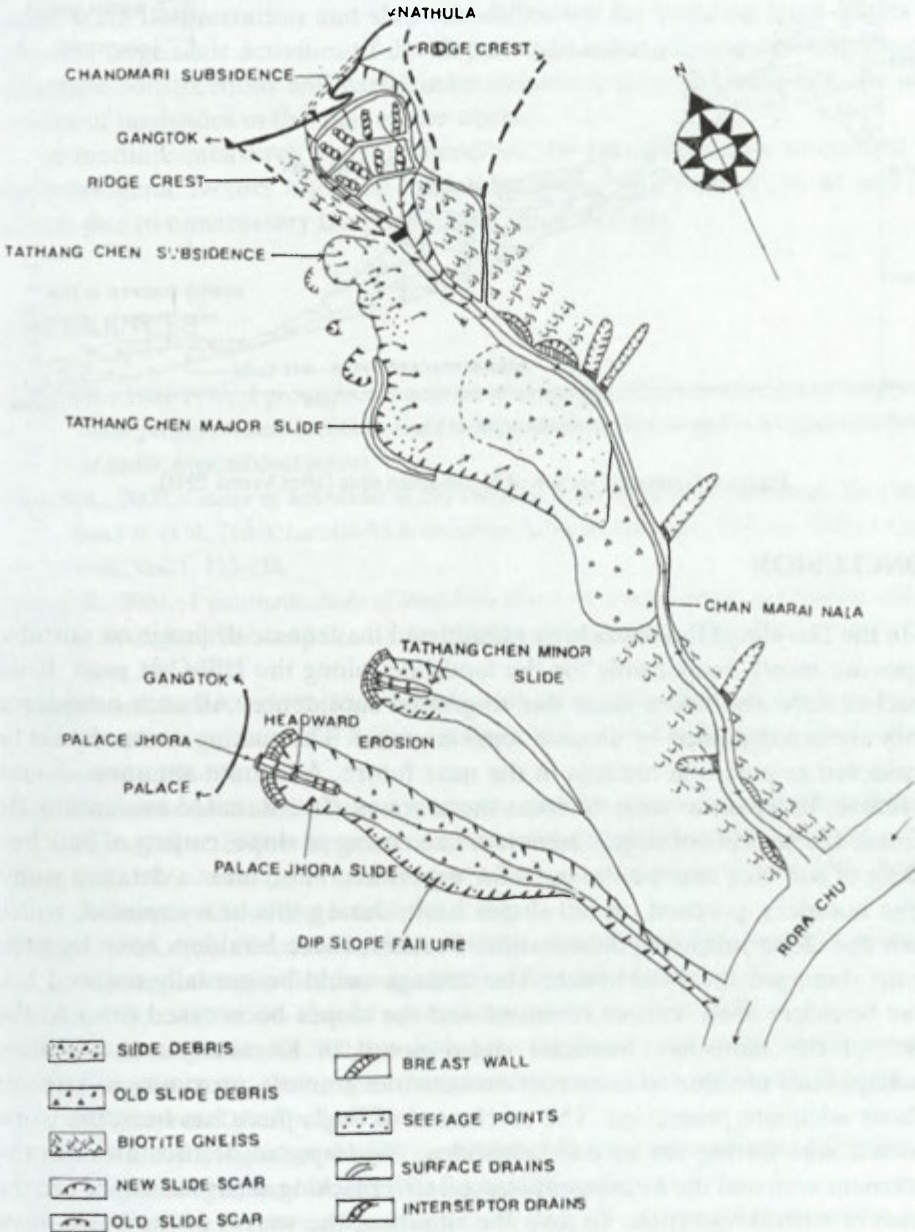


Figure 5. Plan of Chandmari-Tathangchen slide complex (after Verma 1993).

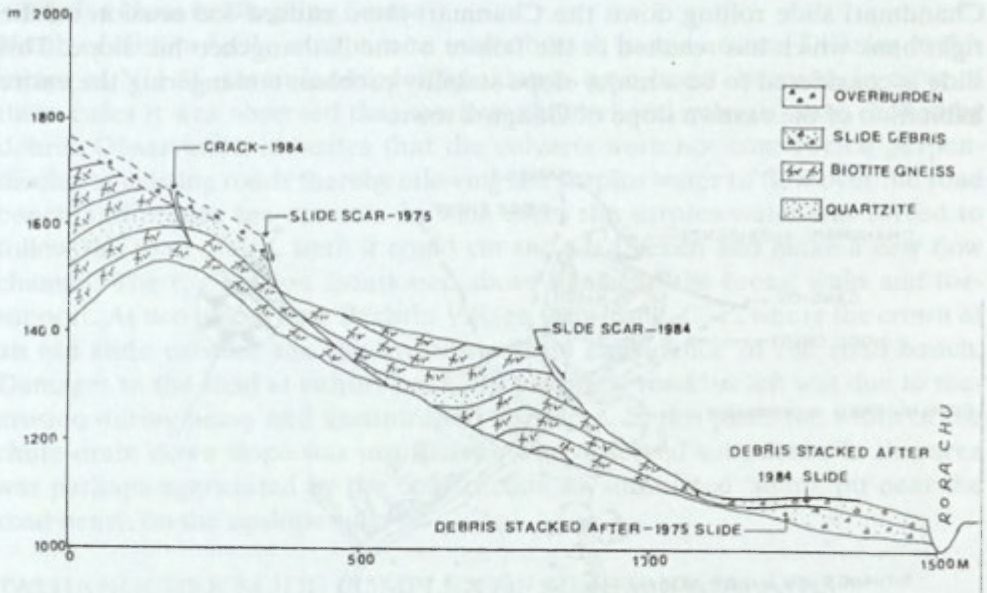


Figure 6. Geological section of Tathangchen slide (after Verma 1993).

CONCLUSION

In the Darjiling Himalayas high rainfall and inadequate drainage on unstable slopes are mostly responsible for the landslides along the Hill-Cart road. Road breaches have also taken place due to ground subsidence. All such subsidence zones are accompanied by arcuate tension cracks. The sinking zones should be considered as potential hazards in the near future. Adequate attention should, therefore, be given to ways to treat these zones after detailed evaluation. To increase the stability of slopes measures like easing of slope, cutting of benches, nailing of soil etc., may be the basis for experimentation after a detailed study. Large boulders, perched on hill slopes have, during this heavy rainfall, rolled down the slope initiating debris slide. Further, these boulders have by their impact damaged the road bench. The damage could be partially reduced had these boulders been earlier removed and the slopes been eased prior to the onset of the monsoon. Surficial slides noted in Kurseong and Darjiling municipalities are due to construction activities in close proximity to streams without adequate protection. The catchment of Pagla *jhora* has been the worst affected area during the recent landslides. Widespread deforestation in the catchment area and the heterogeneous soil cover lacking shear resistance are the causes of such devastation. To save the situation, the waters of the Pagla *jhora* should be controlled through a conduit to produce hydro-electricity or to divert the Hill-Cart road to a new direction.

The Gangtok town area shows occurrence of soft phyllite and schists along with gneiss and some mixed categories of rocks having brittle nature. The soil genetic and physiographic conditions suggest the landslide susceptibility of different terrain units depending on their soil association and parent rock properties (Krishna et al., 2000). Heavy rainfall accompanied by cloud bursts, large scale deforestation and slope modification for building huge hotels for tourists, large scale activities of development in terms of road and other super-structure constructions and earthquake tremors are considered to be the main causes of landslides in this part of the world.

Immediate measures should, therefore, be taken specially to control the anthropogenic factors affecting landslides along with restoration of unstable slopes due to unnecessary tampering with slope stability.

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CLIMATIC AND HUMAN IMPACT ON EPISODIC ALLUVIATION IN SMALL MOUNTAIN VALLEYS, THE SUDETES

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ABSTRACT: The Sudetes are a mid-mountain range located in the temperate climatic zone of Central Europe. Deep valley sides are covered with a thick mantle of periglacial regolith. Migrating cyclones cause intensive rainstorms, debris flow and floods. Geomorphological and sedimentological traces of a previous large flood have been found in the upper course of the Bila Opava and Bela valley floors, which drain the northern slope of the Hruby Jeseník massif, 1000-1400 m a. s. l. Dendrochronological investigation has shown that this large flood took place around the turn of the 20th century. Meteorological archival records have confirmed that extremely heavy precipitation occurred here on 9 July 1903. The periglacial regolith covering the steep, deforested slopes were the source of the coarse-grained clastic material supplied into the Bila Opava and Bela river beds.

KEY WORDS: climatic and human impact, mountain valleys, alluviation, braided river pattern, Sudetes.

LOCATION AND PROBLEM

The Sudetes are the mid-mountain range located in the temperate climatic zone of Central Europe. The average elevation of the Hruby Jeseník massif, located within Eastern Sudetes, is 1000-1400 m a.s.l. (Figure 1). The Hruby Jeseník massif consists primarily of Devonian orthogneiss, locally of migmatites and in the upper part fine and medium-grained paragneiss (Sawicki 1995). During the cold Pleistocene periods, the northern foothills of the Sudetes were probably twice reached by the Scandinavian ice sheet. During the last glaciation (the Vistulian) the ice sheet margin reached within around 200 km north of Hruby Jeseník massif. This created conditions conducive to intensive physical weather-

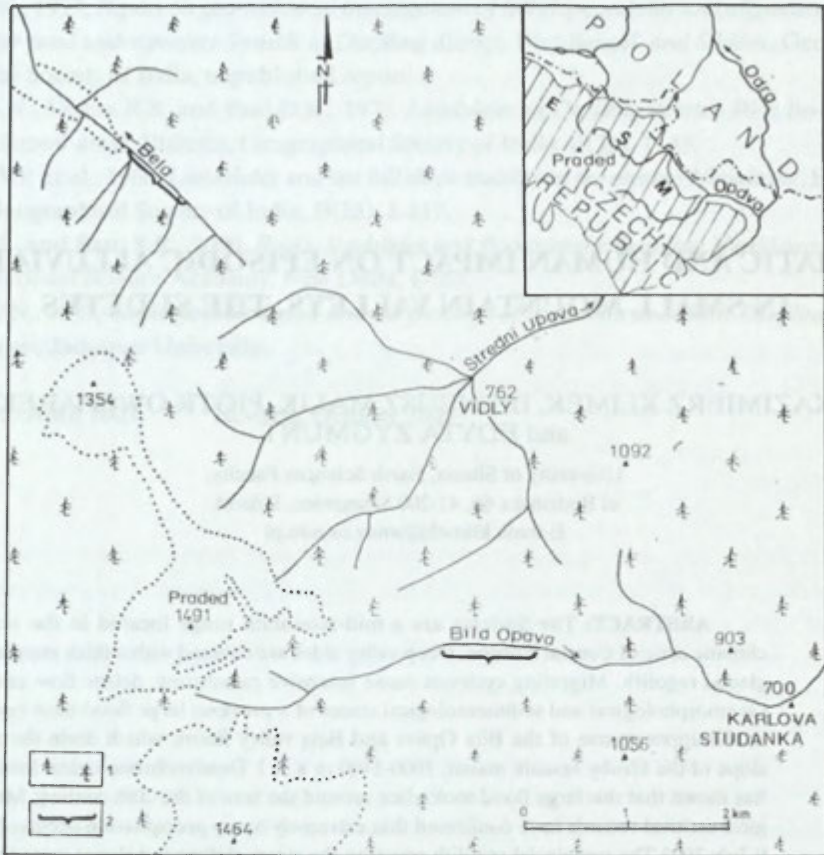


Figure 1. The study area in the Eastern Sudetes Mountains: 1-forested area, 2-valley sections under examination, ESM-Eastern Sudetes Mountains.

ring. Thick regolith covers of varied lithological structure and age were formed during the cold Pleistocene periods (Czudek 1997, Klimek 2000). Depending on the local variations in the lithology of the bedrock and on their position on the valley slopes, these covers may include blocks of over two metres in diameter. In other places, usually in the lower sections of less steeply inclined slopes, these are mixed clay and block covers up to several metres thick, prone to liquefaction when excessively saturated with water, particularly after heavy precipitation (Photo1).

The accumulation of coarse-grained alluvia in the valley floors (the Bila Opaava and the Bela valleys) dissecting the northeastern slope of the Praded massif (1491 m a. s. l.) exhibit braided river patterns in many places. This indicates the occurrence of large, short-duration floods in the past, which led to the accumulation of these coarse alluvia. This paper attempts to investigate the reasons for these phenomena and the time when they occurred.

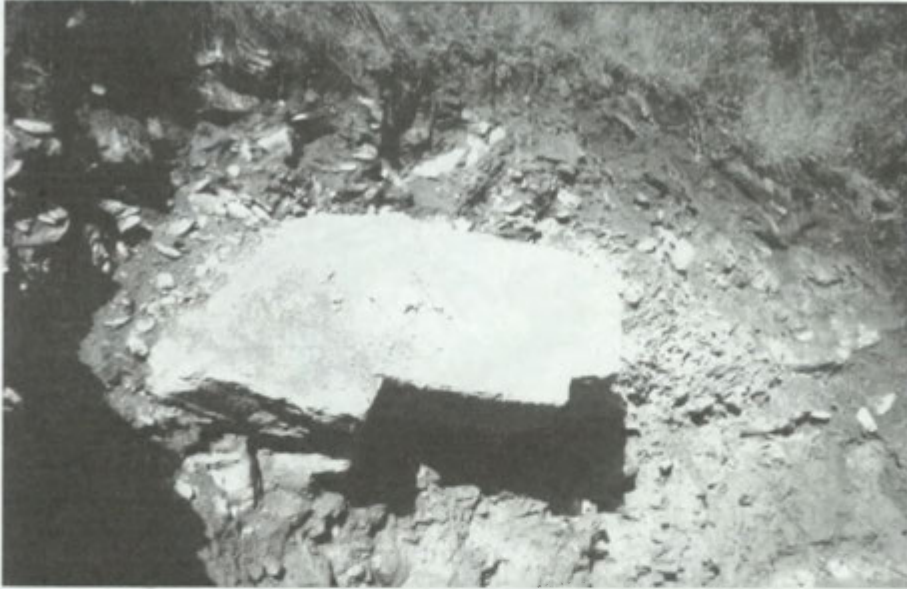


Photo 1. The mix-grained slope covers in the Hruby Jeseník massif.

THE VALLEY FLOOR TOPOGRAPHY AND INFILL

In their upper courses, the Bila Opava and Bela valleys cut to 200-250 m into the northeastern face of the Praded massif. The inclination of their slopes reaches 500 m/km in some places. There is a periglacial slope cover on the valley sides, formed by block fields on the outcrop of quartzitic rock with a fine grained matrix in the gneiss and mica schist zones. There are numerous waterfalls on the quartz vein outcrops in the upper course of the Bila Opava channel. Further down the valleys their longitudinal slope exceeds 60 m/km and the width of the valley floor is 60-80 m.

The exquisitely preserved braided river pattern in the upper course of the Bila Opava and part of the Bela consist of a group of boulder palaeo-bars (Figure 2), the majority of which were formed as longitudinal central bars with their lee slope running diagonally to the valley axis. In many places the edges of palaeo-bars are dissected with overflow channels. Residual boulders with diameter up to 2 m, are often found in these channels and also below the undercuts of their edges (Photo 2).

The undercuts of these coarse alluvia indicate that the palaeo-bars are mainly composed of boulders of up to 1 m in diameter, well imbricated in many places. There are rocky controlled sections along the course of the river channel and there are rapids or small waterfalls in the areas where the quartz outcrops are present. The river-bed is incised up to 2 m below the level of the palaeochannels

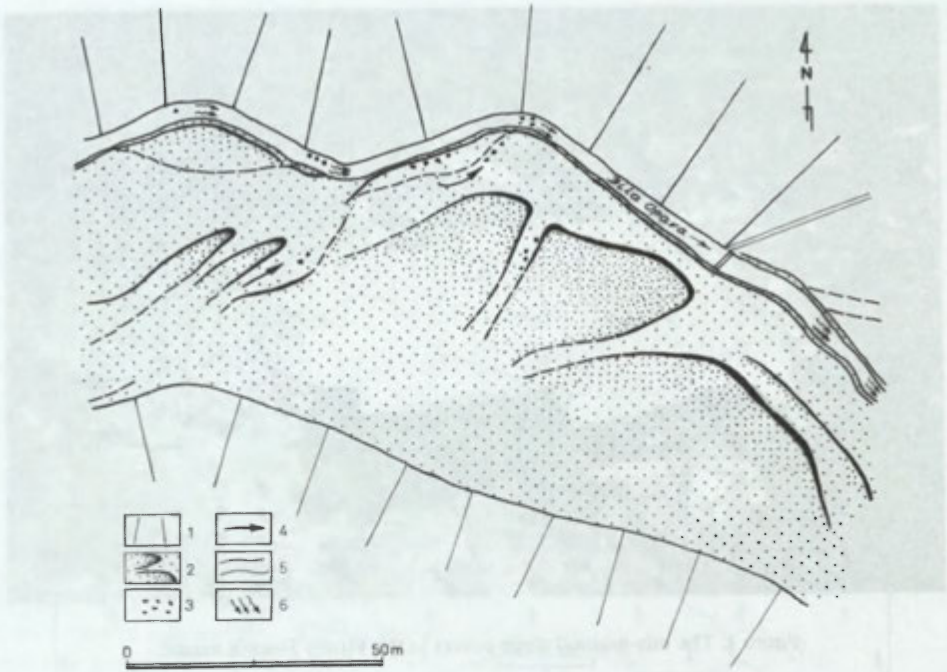


Figure 2. The sub-fossil braided river pattern in the upper course of the Bila Opava valley: 1–valley slopes, 2–sub-fossil braided river pattern, 3–residual boulders, 4–palaeo-flows, 5–present-day channel, 6–rapids.



Photo 2. The residual boulders below the undercut of palaeobar edge in Bila Opava valley.

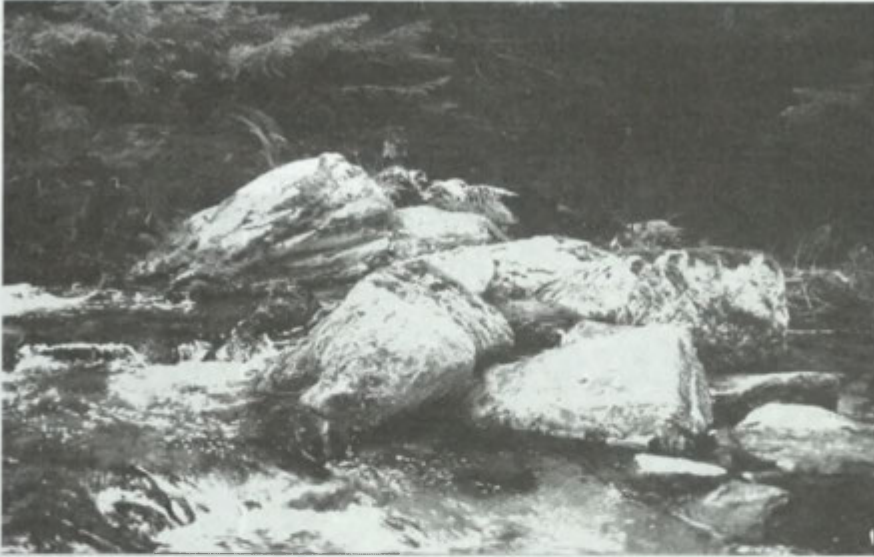


Photo 3. The imbricated boulders in a present-day Bila Opava channel.

separating the palaeo-bars. There are boulders of varying diameter in other places on the bottom of river-bed, the majority of them well imbricated (Photo 3). This indicates the long-term horizontal stability of the river-bed and its tendency to vertical erosion.

The petrographic composition of the clasts reflects the geological structure of the valley sides in this section. The residual boulders of over 1 m in diameter are dominated by quartz, whereas the smaller ones—over 30 cm in diameter—are dominated by quartzite (41 percent) and gneiss (40 percent). The finer clasts are composed primarily of gneiss (up to 60 percent) and quartzites.

The presented fossil braided river pattern in the upper course of the Bila Opava River, as well as the occurrence of similar deposits in the upper course of other rivers in this part of the Hruby Jeseník massif (for example, Bela river) point to the occurrence of very large floods in this part of Eastern Sudetes in the past, with enough energy to transport and deposit quite substantial quantities of coarse debris.

THE SOURCES OF SEDIMENT

The natural upper tree line in the northeastern slope of the Hruby Jeseník massif occurs at the height of about 1300-1400 m a. s. l. It is formed by spruce (*Picea*) forests, but in many places there are mountain pine (*Pinus mughus*) bushes above. There are natural communities of spruce, unaffected by human activity. These natural forest communities usually grow on slopes with block

slope covers. Within these communities the slope surface is usually smooth that indicates no downward movement of slope covers. Pilot dendrochronological research has shown that the trees growing in the upper tree line on these slopes are up to 250 years old. Older trees that die and are periodically felled by natural causes are not utilised for economic purposes.



Photo 4. The spruce tree about 60 years old overgrowing colluvia with older felled stump in Bila Opava valley side.

Tree clearance, which has been practised in the Eastern Sudetes Mountains since the Middle Age, was originally for charcoal production, providing the mining and copper smelters that existed here. Total tree clearance on vast slope areas has probably been practised here at last (?) since the 18/19 century. In the lower section of the slopes the artificially introduced, even-aged spruce dominates. This means that until new trees had been planted, there was no compact forest over large areas of the slopes.

As has been stated, the slopes of the Hruby Jeseník massif are covered by a mantle of periglacial regolith. Some of them, especially the ones located on mica schist outcrops, are prone to liquefy when excessively saturated with water. Debris flows and related phenomena known as 'mura' are generated in such circumstances, especially on the steep artificially deforested slopes of deeply incised valleys. These debris flows in the Hruby Jeseník massif can be over 100 metres wide, 800 metres long and several metres thick (Gaba 1992). There are distinct traces of niches or debris flow tracks on the steep slopes of valleys dissecting the northeastern slope of Hruby Jeseník massif. After the major flood of 1997, many new landslides or debris flows were observed in the lower sections of the valley sides; there were active for several years afterward.

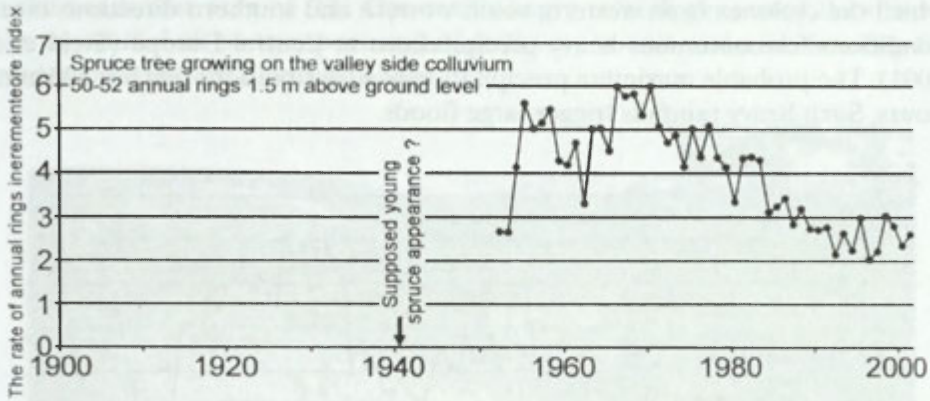


Figure 3. The tree-ring grown increment (1.5 m above ground level) of spruce overgrowing the debris flow niche on the Bila Opava steep valley side.

Detailed research of one of the niches created by mud-debris flow on the steep valley side of the Bila Opava above Karlova Studanka has indicated that tree trunks, which had been felled earlier and were already partly decomposed (the position of the root mass mixed with clay were preserved), were overgrown with even-aged spruce forest (Photo 4). Tree growth analysis conducted at the height of 1.5 m above the ground level indicated 50-52 annual rings (Figure 3). Taking into account the time required for the succession of forest communities and the time required for the spruce to grow to the height of 1.5 metres, we can deduce that the mud-debris flow occurred here around 1940, probably after heavy precipitation on 1 September 1938 or 18-19 May 1940, when the gauge station Vidly located 2.5 km to the north recorded 122.7 mm of precipitation (Polach and Gaba 1998). It is obvious that the processes of slope cover supply into the upper reaches of the Bila Opava and the Bela were the most important, and in many places the only source of point supply of bed load into the river channels cutting through the northeastern slope of Hruby Jeseník massif.

Sediment supplied from the valley sides into the river channels caused the alluviation of the valley floor in headwater areas. In the case of very heavy precipitation and the resulting large quantities of running water, these streams were able to selectively ‘sort’ and redeposit the material supplied from the valley sides. Such a situation, caused by extreme precipitation, occurred in the past in the upper reach of Bila Opava Valley above Karlova Studanka.

THE AGE OF EXTREME FLOOD

The Hruby Jeseník massif receives 1500-2000 mm of precipitation per year. The major part of the intensive precipitation is linked to synoptic situations, in

which the cyclones from western, south-western and southern directions create conditions for continuous heavy precipitations in Central Europe (Stekl et al. 2001). The probable maximum precipitation in such situations reaches 300 mm/2 hours. Such heavy rainfalls trigger large floods.

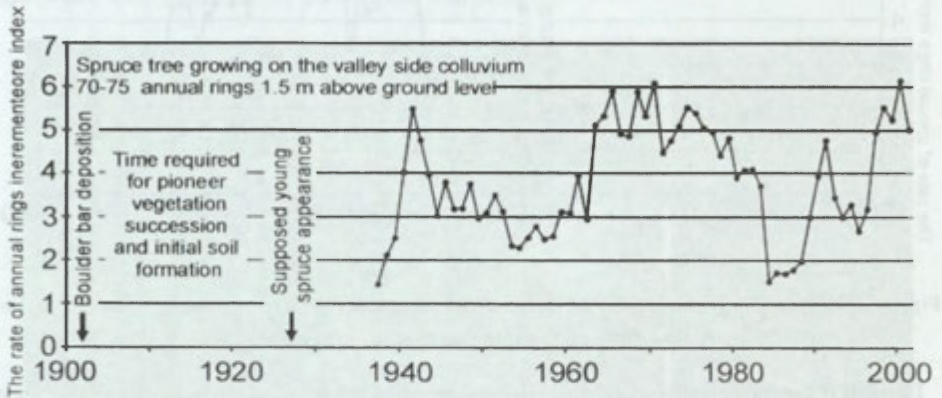


Figure 4. The tree-ring grown increment (1.5 m above ground level) of spruce overgrowing the palaeo-bar of Bila Opava valley: 1-supposed young spruce germination, 2-boulder bar deposition/flood occurrence.

Floods in this part of the Eastern Sudetes were mentioned as early as the 15th century (Polach and Gaba 1998). Records concerning the quantity and intensity of precipitation have existed since 1889 (Stekl et al. 2001). Tree-ring growth increment analysis indicates that the oldest spruce growing on the braided river pattern in the upper course of the Bila Opava valley may be more than 75 years old. Taking into account the time required for the formation of the initial soil and the succession of pioneer vegetation, the spruce succeeded here around the turn of the 20th century (Figure 4). This suggests that the large flood which causes the braided river pattern formation in the valley section under investigation occurred not later than at the beginning of the 20th century. Meteorological sources indicate that, during this period, heavy precipitation in the region of the Hruby Jeseník massif occurred in 1897, 1899, 1900, 1902, 1903, and 1906 (Polach and Gaba 1998, Stekl et al. 2001). The heaviest precipitation, which caused floods throughout all of Central Europe, occurred in July 1903. On 9 July 1903 monitoring stations situated within Hruby Jeseník massif, at the height of 310-775 m a. s. l. received over 200 mm of precipitation: Nova Cervena Voda –240.2 mm, Rejviz –221.0 mm, Sumny Potok –217.7 mm. The hilltops of the massif, located at 1200-1400 m a. s. l., which form the headwater area of the Bila Opava and Bela river tributaries, could have received much more precipitation. This caused cataclysmic floods in the upper reaches of the rivers flowing from the area, which were mentioned in many local chronicles and publications (Palach and Gaba 1998).

In the headwater area of Bila Opava this heavy rainfall generated flood wave that was sufficient to transport and deposit very coarse material, forming the braided river pattern discussed above.



Photo 5. The view of Bila Opava valley from the beginning of the 20th century; note the Praded on the background and almost deforested slopes.

At that time, the supply of silty-debris sediment from slope covers into the river channels was facilitated by the fact that the slopes were deforested to a large degree than is the case now. The view of the Bila Opava Valley with Praded (1492 m a. s. l.) in the background, originating probably from the beginning of the 20th century (Photo 5) but included in a later publication (Grossdeutschland ...1939), shows almost totally deforested slopes. This indicates the possibility of a large supply of coarse-grained slope material into the river channels.

DISCUSSION AND CONCLUSIONS

In the mid-mountain zones of Central Europe, where coarse-grained slope covers were formed during the cold periods of the Pleistocene, debris flows, landslides and other modes of downward slope cover movements are common phenomena. The largest dynamics with regard to these movements may be observed on the steep slopes of Karkonosze massif, above the tree line (Migoń et al. 2002 and the sources cited there). According to these authors, debris flows and related mass movements occur on a smaller scale within other massifs of the Sudetes, e.g., within the Hruby Jeseník massif. The research results presented concerning the alluviation in small valleys cutting through the north-eastern slope of this massif, particularly within the headwater area of the Bila Opava, indicate that the slopes in this area were very active in the past and this phenomenon cau-

sed the local alluviation of small mountain valleys. This intensive slope activity resulted from the exposure of periglacial slope covers due to the intensive exploitation of forests growing on the steep valley sides. Extremely heavy precipitation created the conditions for the liquefaction and movement of these slope covers towards valley floors. The current avoidance of large slope clearance now limits sporadic debris flows to small areas.

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HIGH-MOUNTAIN VALLEY FLOORS EVOLUTION DURING RECESSION OF ALPINE GLACIERS IN THE MASSIF DES ÉCRINS, FRANCE

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ABSTRACT: The age of recent glacial, glacialfluvial/fluvial forms is determined using lichenometric dating in two Alpine valleys of the Massif des Ecrins in France. A lichen growth-curve is based on data from natural sites, such as boulder-fields, moraine ridges etc. of known age on recently deglaciated terrain. The maximum extent of the glacial system of the Venéon and Etançons valleys during the Little Ice Age was 15 km in length. The maximum lichen diameter of 90-95 mm characterizes the stage of recession correlated with AD. 1650-1660. Analysis of the systems of terraces and paleochannels of the proglacial rivers supported by lichenometric dating allows three periods of intensified fluvial activity to be distinguished for the Little Ice Age.

KEY WORDS: deglaciation landforms, lichenometry, Little Ice Age, Massif des Ecrins, France.

INTRODUCTION

In the period 1998- 2001 field studies were carried out in the French Alps by the Polish-French geomorphologic group as part of the Polish Academy of Sciences (PAS) and Centre National de la Recherche Scientifique (CNRS) exchange programme. The aim of these studies was to establish geomorphologic regulari-

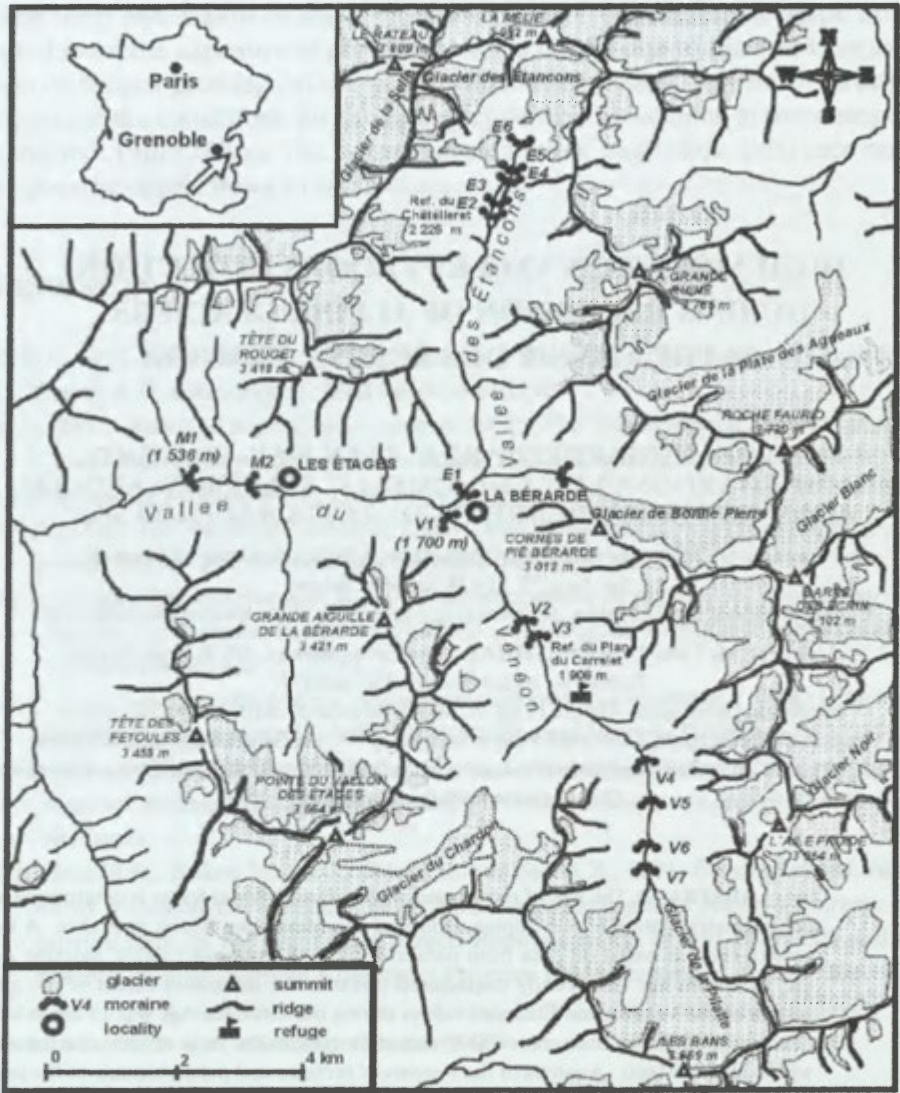


Figure 1. Sketch map showing the location of main study areas in the Écrins massif, French Alps.

ties in valley floor evolution during Little Ice Age (LIA) deglaciation in the Massif des Écrins (Barre des Écrins 4102 m a.s.l., La Meije 3982 m; N 44°, E 6°). Reconstruction of past geomorphic events was possible by using lichenometry for dating landforms and sediments in the foreland of melting glaciers. The glacialfluvial, fluvial and slope processes were conditioned by climatic changes. The complexes of forms such as moraine ridges, glacialfluvial fans, alluvial/glacialfluvial terraces in the bottoms of the main valleys, as well as lateral glacialfluvial cones in

front of tributary hanging glaciated valleys, gravitation and debris flow cones, and talus slopes produced by rockfalls/rockslides are very distinct in such environments. The main research concentrated in the area of the Vénéon and Étançons valleys with some additional, comparative work in the La Selle valley and Ailefroide valley.

The lichenometric method has been broadly applied in dating LIA and more recent landforms. It was used in numerous studies of rapid mass movement both in alpine and arctic environments (e.g., Blijenberg 1998; Helsen, Koop and Van Steijn 2002; Jomelli 1999; Lewkowicz and Hartshorn 1998). There are also attempts to use this method in the study of landform evolution in proglacial areas (e.g., Gregory 1987, Gordon and Sharp 1983).

In 1998 and 1999 the most important aim of the field studies was to construct a lichen growth curve for the Massif des Ecrins. The aggregated lichen species *Rhizocarpon*, one of the more commonly used groups in lichenometry, was employed. Old gravestones in cemeteries in the St.Christophe en Oisans and Venosc villages, ruins of old stone buildings and rockslides of known age gave the first indications of a lichen growth rate in the main valley of Vénéon. Lichenometric data collected on moraine ridges dated according to cartographic sources, historical notes and old photographs made it possible to construct another lichen growth curve (Baumgart-Kotarba et al. 2003). This second curve is based on data from natural sites. It is characterised by a faster lichen growth and is more appropriate for dating landforms in highly elevated areas, more opened to solar radiation.

The maximum extent of the glacial system of the Vénéon and Étançons valleys during LIA was 15 km in length with the maximum frontal moraine located at an altitude 1536 m, down valley from Les Étages village (Figure 1). Morphological evidence of the separation of the Vénéon and Étançons glaciers is found close to La Béarde village. Étançons glacier was hanging above La Béarde at 1760 m a.s.l. (E1 in Figure 1), and the Vénéon glacier in the valley at 1700 m a.s.l. (V 1 in Figure 1). Maximum lichen diameters of 90-95 mm (mean max* 92.5 mm) characterise this stage of recession and could be correlated with the years 1640-1660.

THE VÉNEON AND ÉTANÇONS VALLEYS

The upper reach of the Vénéon valley, from Les Bans summit (3669 m) to the north to La Béarde village (1700 m) is 10.7 km long. The Vénéon torrent is supplied with water by many tributary valleys filled with glaciers. In longitudinal profile one can distinguish several steps. The highest steps represent the Glacier de la Pilatte with distinct levels: at 2700 m and 2600- 2550 m. A recent position of

* Mean max denotes the mean of the five largest measured diameters.

the glacier front is located at an altitude of c. 2250 m. From 2200 to 2120 m an active glacialfluvial fan has slopes of 17.7 percent. Between 2120 and 2070 m the first extraglacial alluvial/glacialfluvial braided plain has been developing inside the moraine arc since AD. 1850. Between 2070 and 2040 m the Vénéon torrent is incised in the moraine gorge with a gradient of 10 percent. An alluvial reach located upstream has a gradient of 7.1 percent and the next downstream alluvial reach slopes at 5.7 percent to an altitude of 2000 m. The next reach is located at altitudes from 1970 to 1870 m (slope 4.2 percent) (Carrelet). From 1870 to 1840 m there is a steeper, 15 percent reach, followed by a gently sloping (5.5 percent) stream reach is found at altitude 1840-1790 m (Pierre Chamoissiere). A steep reach with the gradient of 10.7 percent divides the next alluvial reach (La Béarde) being 1 km long with the gradient of 4.3 percent (Figure 2).

On both sides of the Vénéon valley in La Béarde a blocky moraine is found. On the left side there are two distinct ridges while on the right side only individual granite blocks near the edge of the La Béarde terrace were found with maximum lichen diameter 90 mm. The present riverbed is incised c. 10-15 m below the glacialfluvial terrace of La Béarde. Below the village it is possible to distinguish two widenings of the valley bottom; at Les Étages (1595 m) and Pièce du Clot (1536 m). There are two terminal depressions related to the maximum (M1) extent of the Vénéon Glacier—a 15 km long glacier, and to the first recessional stage (M2)—a 13.4 km long glacier (Figure 3).

The Étançons valley culminates on the summit of La Meije (3982 m) and is opened to the south. The outlet of the valley hangs 150 m above the bottom of the Vénéon valley in La Béarde. The recent extent of the Étançons glacier is identified at an altitude of c. 2700 m. The foreland of two glaciers, Etançons (NW) and Pavé (NE), in the upper part consists of the steep sloping zone (27 percent) from 2700m to 2320 m a.s.l. and further down – a gentler part (15.5 percent) (Figure 4). The second part comprises the area between the 'Conic' moraine ridge E4 (2320 m) and the double glacialfluvial fan of Grande Ruine glacier system (2150 m). In its lower part recessional moraines E2 and E3 close to Refuge du Chatelleret are preserved. E3 moraines are covered by *Rhizocarpon* lichen 55-48 mm in diameter (mean max 50.5 mm). The next reach can be distinguished at altitudes from 2150 to 1870 m. A lower, 3.8 km long, part of the valley bottom has a gradient of 7.3 percent to the mouth of the Bonne Pierre valley. Thus, this flat reach is morphologically comparable with the Carrelet reach in the Vénéon valley. The rocky step of the hanging Étançons valley is deeply dissected a by gorge. On both sides of the gorge, the lateral moraine ridges persist. On the left side, there is the moraine sloping from 1860 to 1840 m where *Rhizocarpon* lichen 90-95 mm in diameter have been identified. On the opposite side, there are two lateral moraines. The lower one, used partly as a tourist trail, has yellowish weathered fine grain material at the height 1840-1860 m. The upper one of unknown age can be identified at the height c. 1940 m where it forms a debris shelf covered by an alpine meadow. The left lateral blocky moraine ridge is very

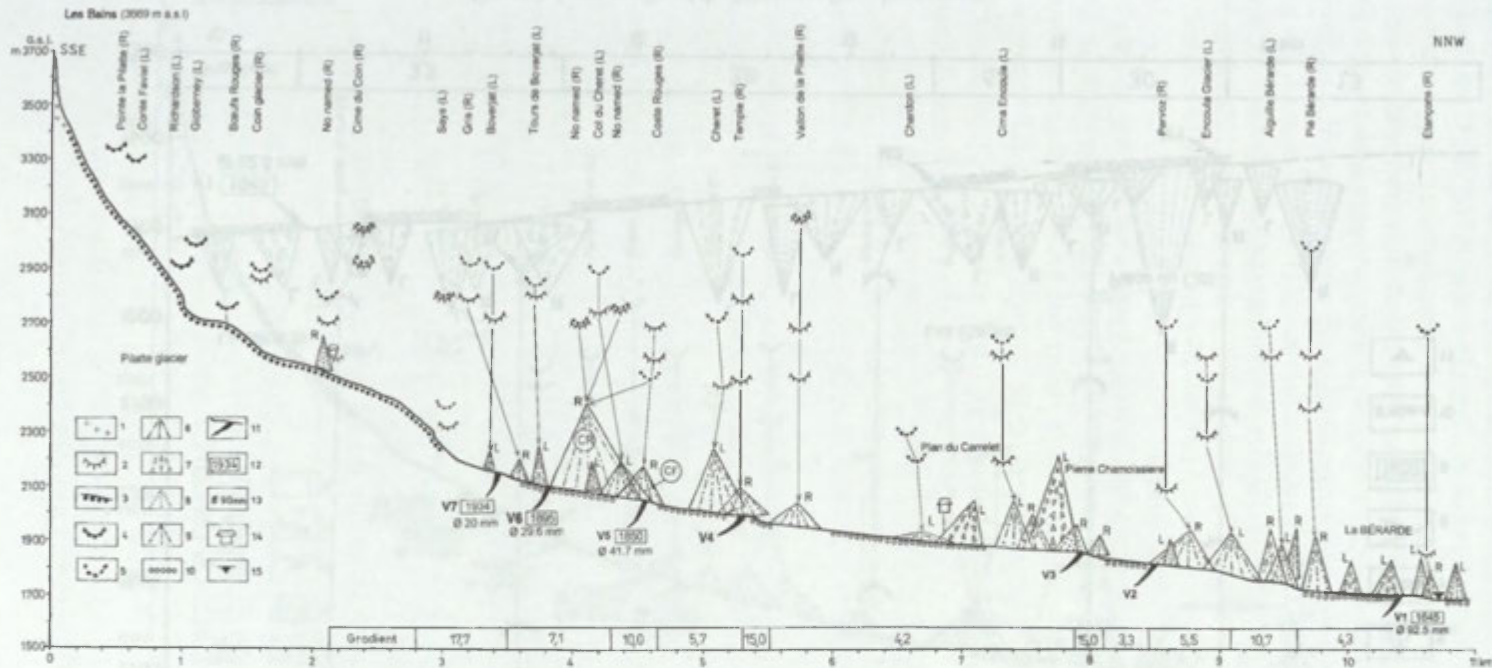


Figure 2. Long-profile of the Vénéon valley. 1 – crystalline bedrock (rowall), 2 – step of hanging tributary valley, 3 – recent surface of main glacier, 4 – recent hanging glacier, 5 – front of glacier in hanging tributary valley, 6 – glaciifluvial cone, 7 – rockfall/rockslide cone, 8 – talus cone, 9 – alluvial talus cone, 10 – glaciifluvial/fluvial plain in valley floor, 11 – recessional moraine stage dated by lichenometry, 12 – lichenometric data (AD), 13 – *Rhizocarpon* mean max lichen thalli diameter in mm, 14 – mountain refuge (challet), 15 – confluence of Étançons and Vénéon torrents, L – left side landform, R – right side landform. gradient in %, CR-Coste Rouge cone, CF-Cloute Favet.

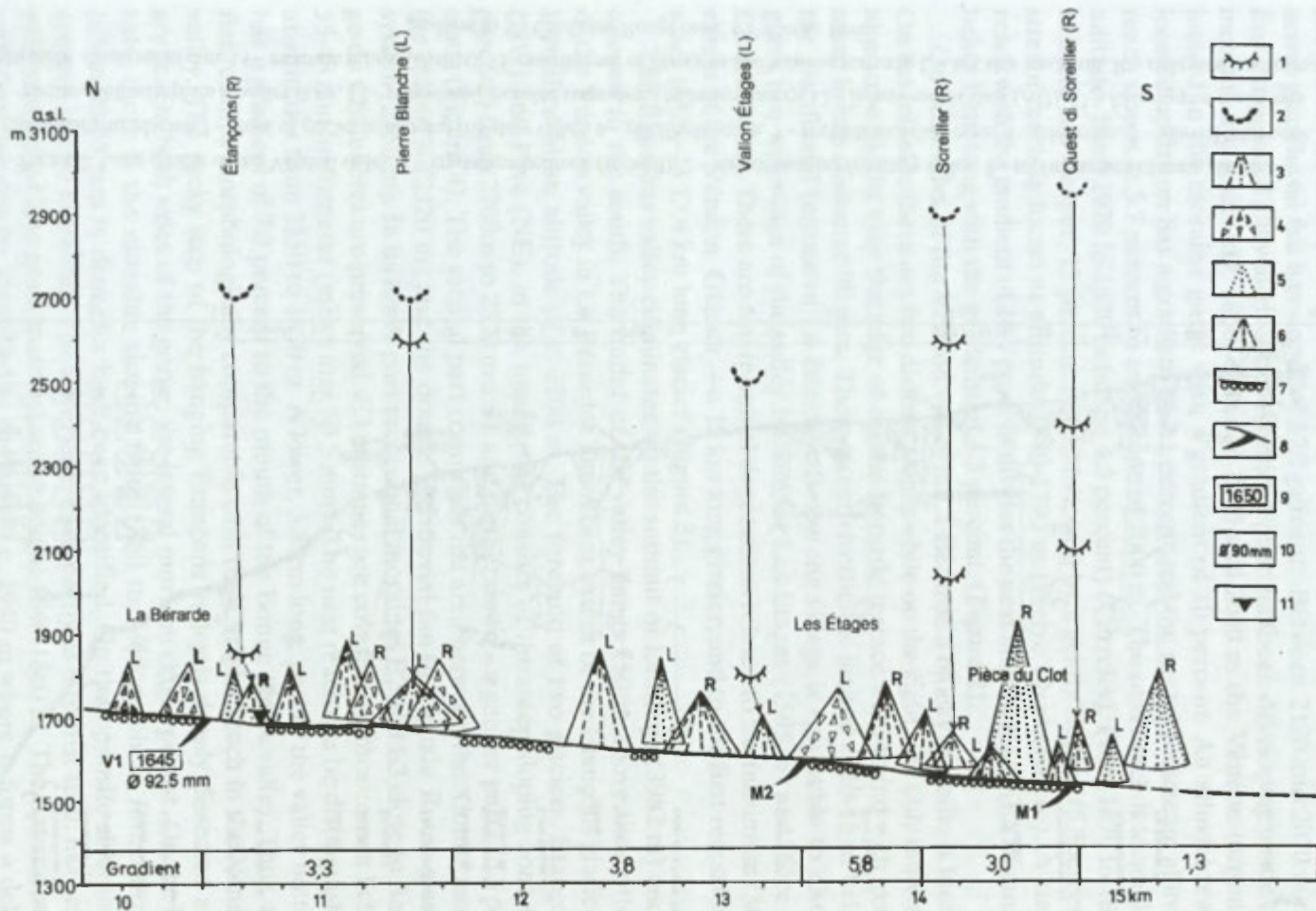


Figure 3. Long-profile of the Veneon valley – lower reach.

For explanation to symbols see Figure 2.

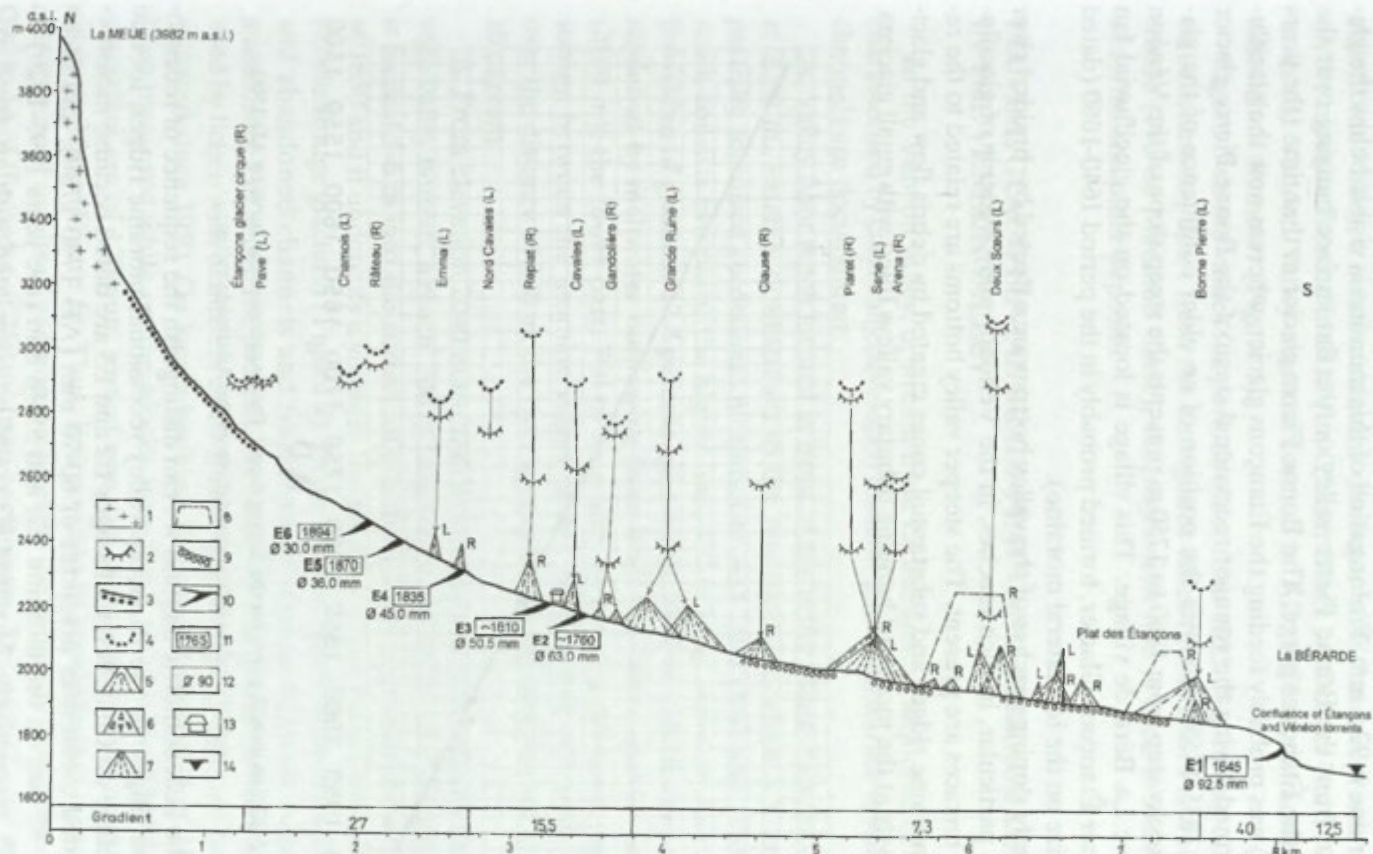


Figure 4. Long-profile of the Etançons Vénéon valley.

For explanation to symbols see Figure 2.

well preserved on the Plat des Étançons. The mean maximum diameter of lichen on this moraine is 92.5 mm. Prolongation of this moraine is visible below the glacifluvial fan from the Bonne Pierre valley on the flat surface hanging over the incision of the Étançons gorge. The Bonne Pierre glacier at that time (the years 1640-1660) was probably feeding the Étançons glacier, whereas now the glacifluvial fan formed during the younger recessional stage of the Bonne Pierre glacier (the years 1835-1855) occupies the position of an older confluence of the glaciers. The rocky step from 1860 to 1750 m presents the steep slope of the Vénéon valley above La Bérarde village. This village is located on the glacifluvial fan of the former Étançons Glacier, formed probably in the period 1640-1660 (dated by lichen size on the left lateral moraine).

The gently sloping reaches of the valley bottom are affected by braided river activity. In particular, in such reaches, in the Vénéon valley, systems of glacifluvial/fluvial terraces are present. The steeper valley bottoms are related to the recessional moraine ridges and substantial cones created by debris flow and glacifluvial activity at the mouth of hanging tributary valleys filled with small glaciers.

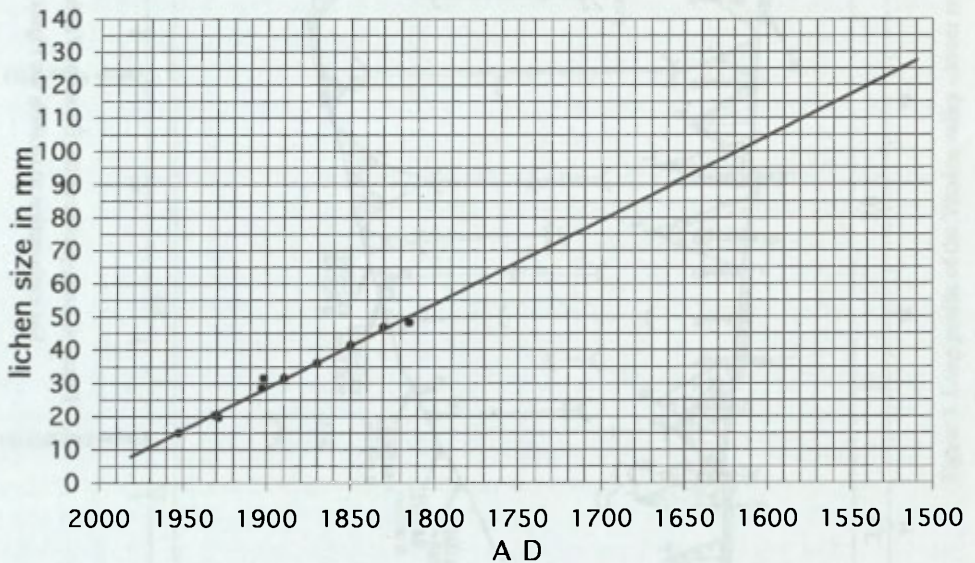


Figure 5. Lichen growth curve for the Écrins massif. Prolongation of the curve to AD 1500, partly based on lichen size data from the Ailefroide valley.

Using the lichenometric method we can distinguish the sequence of recessional moraine ridges. In the Étançons valley recessional moraine ridges, located close to Refuge du Châtélleret, of stages E2 and E3 are dated because measurements yielded mean max diameter of 60-70 mm (AD 1740-1780), and mean diameter – 63 mm on the moraine E2, and 55-48 mm (the years 1795-1820) on moraine E3, respectively. Moraine E3 could be correlated with an age c. AD 1810 by comparison with the age of a known recessional moraine located close to

Refuge Cezanne in the Ailefroide valley. According to our measurements in the Ailefroide Valley, a mean maximum diameter of 48.7 mm represents lichen size on the moraine formed in 1815 (Le Glacier Blanc Le Glacier Noir, 1995), whereas in the Vénéon Valley moraine formed in AD. 1850 is characterised by 41.7 mm lichen diameter (Figure 5).

Figure 5 shows values for periods older than 1815. Not having older reference benchmarks, apart from the ambiguous dates of maximum moraines in Ailefroide (Pré de Madame Carle), the regression line is used to estimate an age of forms and older deposits.

FLUVIAL/GLACIFLUVIAL BRAID-PLAIN AND TERRACE LEVELS IN THE VÉNEON VALLEY

In the Vénéon valley, upstream from La Bérarde, a sequence of five reaches characterised by classically developed braided-plain and terrace levels with paleochannels are documented:

A. Valley floor reach located in front of the active Pilatte Glacier, now located at 2250 m., extends downstream to the recessional moraine V5, and overrides partially fossilised recessional moraines of AD 1895 (V6) and 1934 (V7). This reach consists two parts: the higher one, 450 m long, represents an active glaci-fluvial fan of gradient 17.7 percent, and width 50-70 m in its lower section and is modelled by meltwater outflowing from both the Pillate and La Says Glaciers (2180 m); the lower part, 700 m long and 7.1 percent gradient, represents the section between the moraine ridges of AD 1895 and of AD 1850. Downward, over the distance of 300 m, the river gorges across the moraine and its gradient is 10 percent.

B. From elevation 2040 m to 2000 m braided channels occupy the 120-180 m wide valley bottom, a reach 700 m long with a gradient of 5.7 percent. The reach is incised 5-6 m into the surface of glaci-fluvial fan associated with the moraines of 1850 and it undercuts a younger the glaci-fluvial fan from Cloute Favier Ravin (Coste Rouge glaciers) (Figure 2). In the alluvial reach there are active channels and abandoned channels and fragments of terraces scarcely overgrown with grass rising 3 m above the present-day channel. The gravel surfaces are characterised by lichen with diameters of 10-15mm, which suggests that they have not been accreted since 1950. On larger boulders lichen with diameters of 110 mm have been measured while on the blocks from the 5 m high glaci-fluvial terrace the diameters are of 120 mm, so that one can assume that these blocks originated from rockfalls (Figure 6). Thus, it is possible to conclude that during 150 years the erosional results of geomorphic work by the proglacial torrent can be estimated to be of the order of 750 m³ from each meter. The next alluvial reach is separated from reach B described above by 200 m long reach sloping at 15 percent.

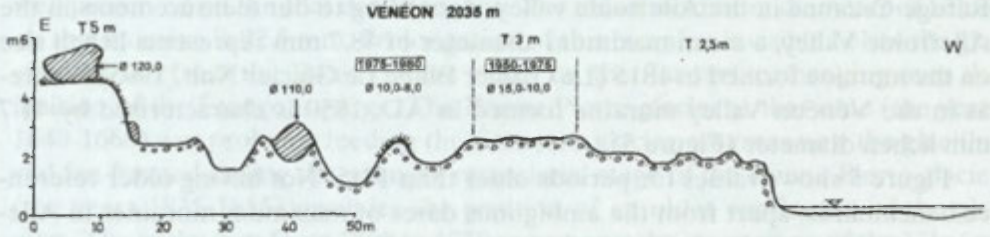


Figure 6. Cross-profile of the Vénéon valley downstream of moraine ridges from AD 1850. Mean lichen diameters for five largest thalli and calculated age.

C. Carrelet, the next reach, delimited by altitudes 1970 and 1870 m, is 2.4 km long and slopes at 4.2 percent. This largest flat part of the bottom of the Vénéon valley is up to 500 m wide. The extent of this reach is probably conditioned by an extensive valley over-deepening at the junction of Chardon and Pillate-Vénéon glaciers. The 1 km long section located upstream of the Carrelet plain has a 50-75 m wide alluvial bottom sloping at 5-6 percent. The widening of the Carrelet plain comprises a vast fan (350 m in radius) of the Chardon stream, active Vénéon channels and a system of terraces of heights 5 m, 3 m and 2 m (Figure 7). The 5-6 m high terrace can be related to the period before 1815, as the rounded pebbles preserved in the paleochannels have lichens of 50-38 mm in diameter (mean diameter 43.5 mm). The extensive level of the undulated plain, forming the 3 m high terrace, can be dated at the years 1870-1890 based on the measured lichen. At the level of the 3 m high terrace an inner reach (nearby slope reach), was dated at 1852-1895 according to lichen, and an outer reach formed in 1875-1895 can be distinguished. Braided rivers formed the 3 m high terrace, so that in the vicinity there is material from bars that accreted during various floods. In general, however, the outer reach of the 3 m high terrace is 20 years younger. Considering the mechanism of terrace formation by the braided rivers, it should be emphasised that the functioning of paleochannels on the 5 m high terrace is associated in time with the oldest channels found on the 3 m high terrace. The 2 m high terrace was formed in the period 1875-1895, i.e., in a similar period to the outer part of the 3 m high terrace, but a mean age of the 2 m high terrace can be estimated as 1880, i.e., 10 years younger than the younger part of the 3 m high terrace. The lowest terrace level, 1m high, has no lichen, because it is a part of the active Torrent Vénéon fluvial system. It should be noted here, that according to a tourist guide of 1854 (Roussillon 1979) the Chardon glacier terminated at an altitude of 2000 m, i.e., it was about 0.5 km distant from the present-day Vénéon channel. The 2 m high terrace below Refuge du Carrelet tourist shelter was formed in 1870-1902-1920 (Figure 7, profile P2). In profile P1 the 1.5-2 m high terrace consists of elements formed in 1888, 1922, 1930, 1916, 1915, which suggest that the mechanism of this terrace formation was also the mechanism of forma-

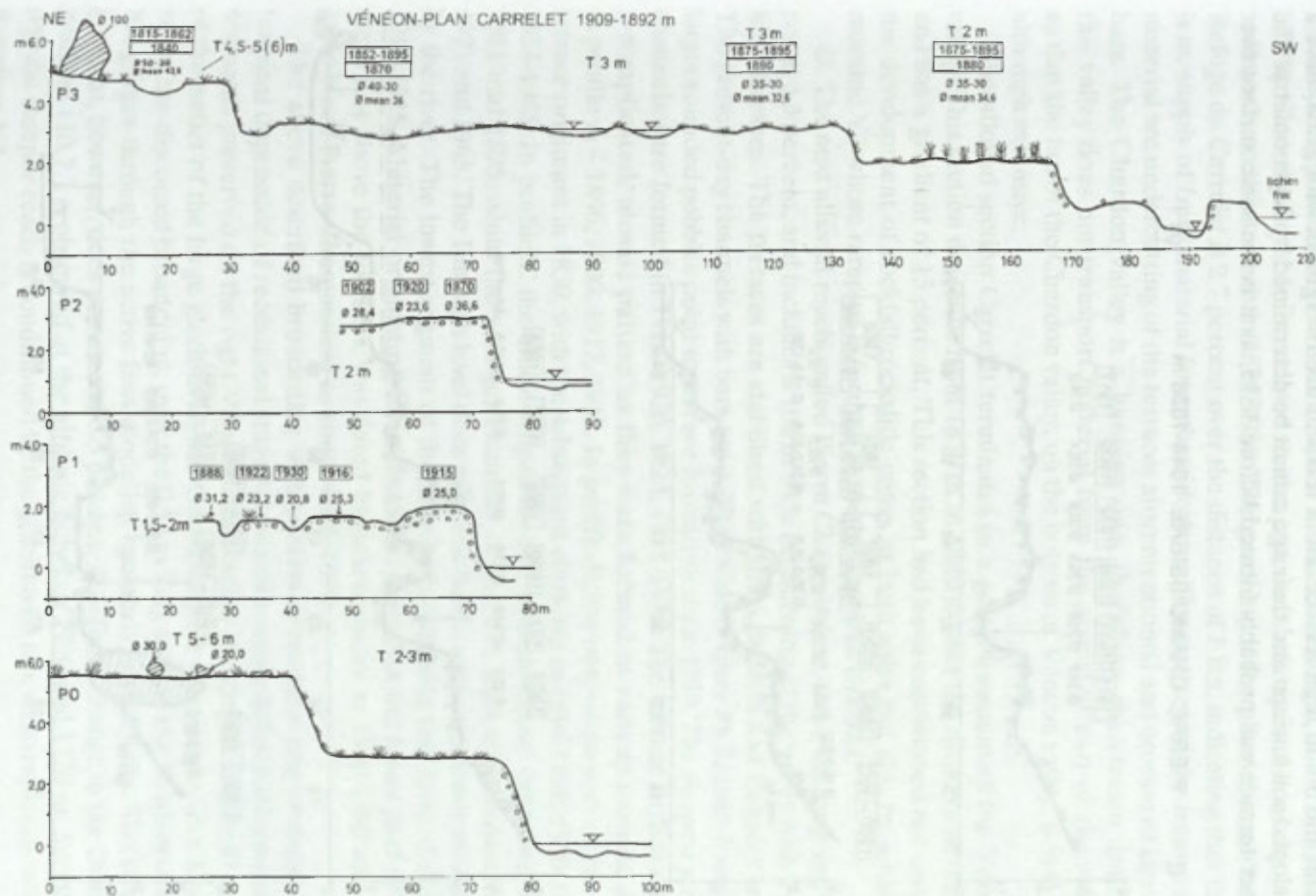


Figure 7. Cross-profiles of the Veneon valley: Carrelet terrace levels.

tion of the braided river deposits. In profile P0 there is only a 2-3 m high terrace besides the 5-6 m high terrace. Both terraces are overgrown with grass, have a developed soil horizon and their age cannot be determined by lichenometry. The higher terrace was probably formed before 1815, as it is the same surface over

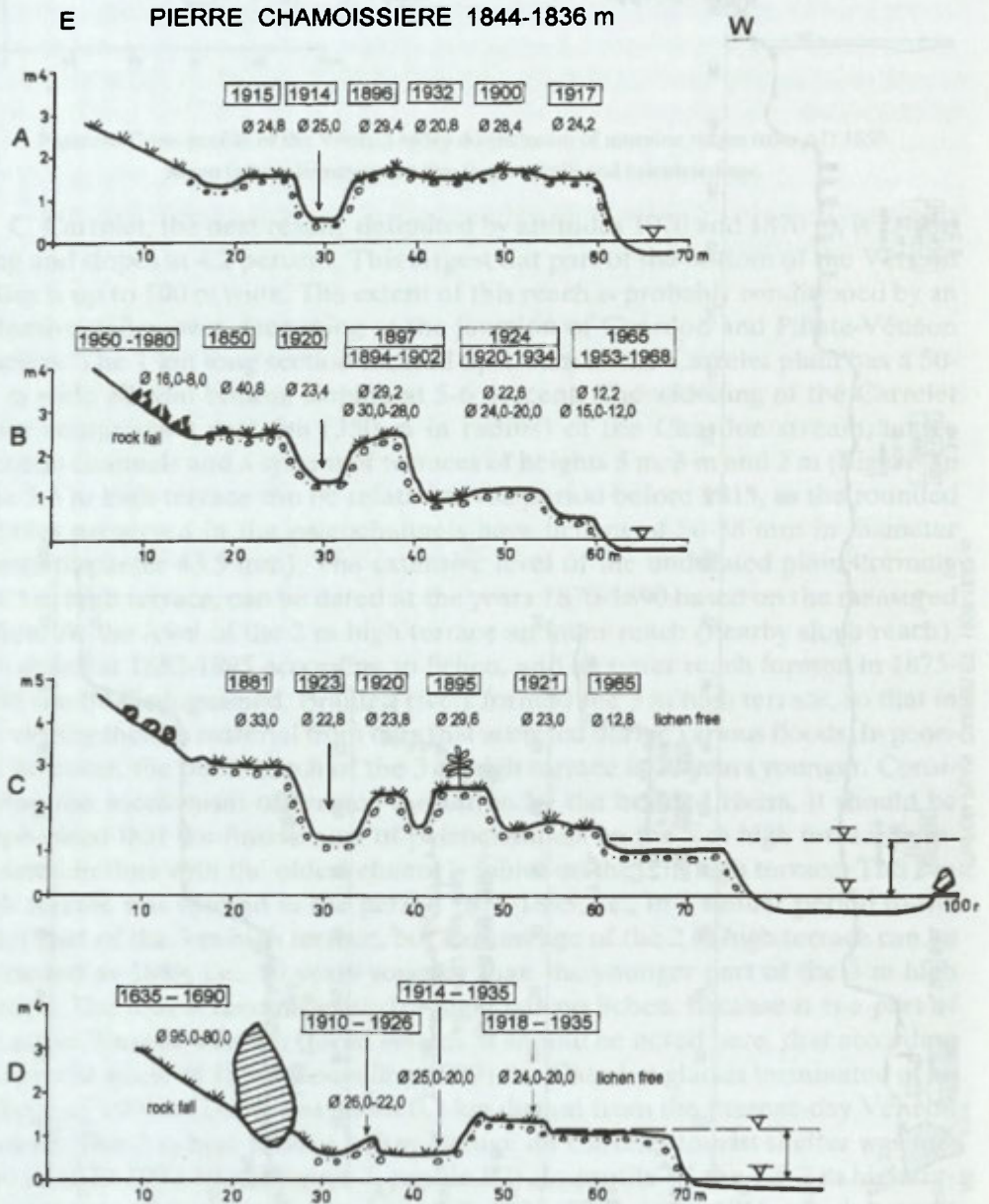


Figure 8. Cross-profiles of the Veneon valley: Pierre Chamoissiere terrace levels.

which the road to Refuge du Carrelet runs. On the surface of this terrace there are blocks deposited by debris-mud flows. The sizes of lichen, 20-30 mm, on the blocks allow the latter to be related to the processes of deposition of the slope sediments on this terrace in 1890-1930. The gradient of the alluvial reach below Refuge du Carrelet is 2.7 percent over the distance of 1 km, indicating that there is no supply of fresh material from the side fans. The sources of the transported material are undercutting of the terraces aforementioned and erosion of channel bars. The Chardon valley is a hanging one, therefore, the stream draining this valley does not transport material from the glaciated part of the valley, so that the fan of the Chardon valley, on the bottom of Vénéon valley, is very flat although extensive.

The alluvial section Carrelet terminates in a gorge fragment of the Vénéon valley. This section descends from 1870 m to 1840 m over the distance of 0.2 km and has a gradient of 15 percent. This section had been conditioned not only by the development of rockfall/rockslide cones at both sides but also by a blocky moraine V3 whose rampart is formed at the left bank of channel.

D. The next alluvial reach, called Pierre Chamossiere, is 0.3 km long and slopes at 3.3 percent, and includes a 100 m wide broadening of the valley with 2-3 m high terraces. The terraces are stabilised with grass, bushes and singular trees. The present-day channels with bars are c. 30 m wide (Figure 8). Lichen found on larger rounded pebbles point to terrace formation since 1850. The deepest paleochannels were formed in 1914, 1920, 1923, 1914-1935. The terrace surfaces have a complicated, mosaic pattern as they were formed in various periods, e.g., in profile A – 1896, 1900, 1917, 1932. In profile B, the near-slope part of the terrace originates in 1850, with the subsequent elements in 1894-1902, 1920-1934, 1953-1968. In profile C the oldest fragments of 3 m high terrace originate from 1881 and 1895, while the younger elements of the 2 m high terrace date from 1921 and 1965. The 1 m high level does not have lichen and is currently modelled by the river. The lowest segment of the alluvial broadening was formed during 1918-1935. Material originating from rockfalls, found on the lower part of the slope just above the terraces, was dated by lichenometry as 1950-1980 and the great block Pierre Chamossiere as about 1635-1690.

The above described broadening with preserved terraces was probably the terminal depression of recessional stage V2, with its extent delimited by moraine remnants preserved on the right Vénéon slope below the cirque threshold south of the outlet of the large glacial fan Grand Clapier (Pervoz glacier). Down-valley the discussed broadening, over the 0.5 km long section the Vénéon stream has gorges through the active fans modelled by debris flow activity. The stream gradient, however, does not exceed 5.5 percent. A definite change in the channel gradient (10.7) is observed at the altitude between 1790 and 1730 m. Supposedly, the steeper reach is conditioned by the glacial fan, which corresponds to moraines V2.

E. The next reach has a gradient of 4.3 percent and occupies the terminal depression of recessional moraines V1 (lichen size 90-95 mm). At present, in this reach there is a parking lot in La Bérarde, which shows that areas of the alluvial plain and the lowest terrace have been artificially levelled. A narrow ledge of 4.5-6 m high terrace has been preserved on both sides of the Vèneon torrent. In Figure 9 fragments of 4.5 and 2.5-3.5 m high terraces are documented.

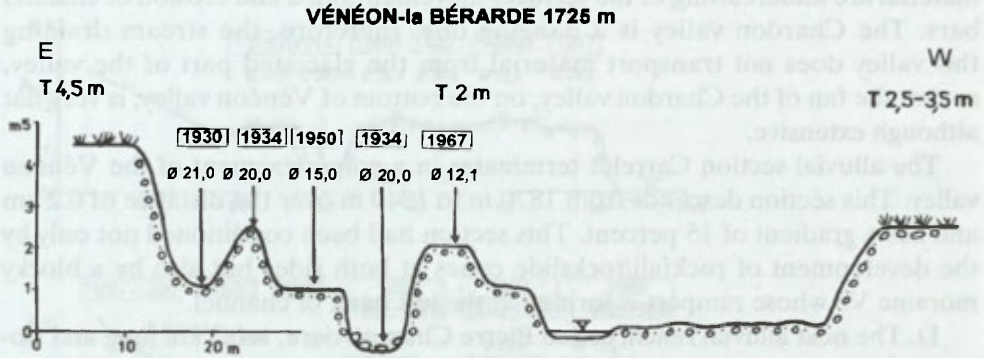


Figure 9. Cross-profile of the Vèneon valley: La Bérarde.

The paleochannels dated by lichenometry at 1930 and 1934 were reconstructed and fragment of bars of 1934, 1950, and 1967 rising 1–2.5 m above the present channel are shown. Generally, this part of the alluvial channels was forming between 1930 and 1967. Thus, this braided plain has formed since 1930, so one could conclude that after 1930 no significant dissection occurred apart from normal braided river activity. Some 60 m below this cross-profile and above the helicopter landing (Helistation) bare granite bedrock is found (Figure 10). Lichen diameters on this granite surface change with distance from, and height above, a

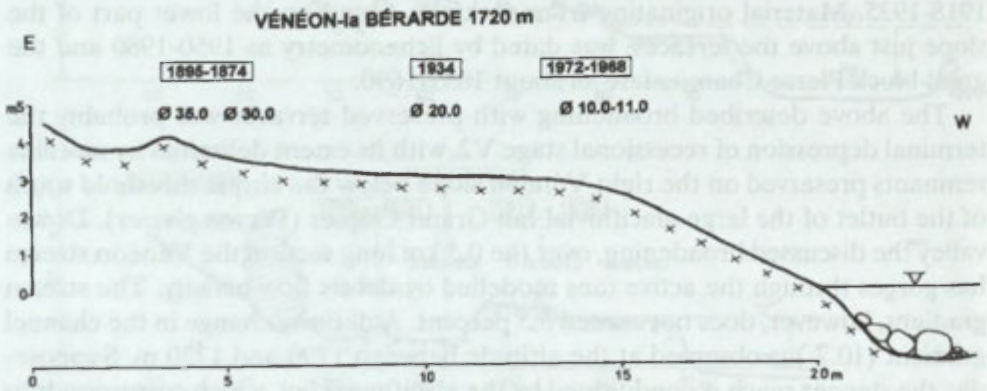


Figure 10. Crystalline bedrock topography in La Bérarde village dated by lichenometry.

river channel. Three metres high valley floor located c. 10 m from the present channel was affected by river activity before AD 1934, while 3.5 m high fragment located 15 m from the channel is dated at the years 1874-1895. A calculated mean rate of fluvial incision of the Torrent Vénéon in the granite bedrock during the last 100 years is 33-43 mm per year. A similar rate of incision has been established in the case of the alluvial 3 m high terrace (37-47 mm/yr), occurring below the glacial polish at Helistation. Determining the rate of incision of the Torrent Vénéon into solid rock is very valuable, as the evaluation of the rate of incision of the covers by braided rivers will remain less precise due to the complexity of modelling channels of high-energy braided rivers.

Down the reach of the Vénéon valley conditioned by the presence of a terminal depression related to recessional moraines V1, two reaches occur: one 1.2 km long and sloping at 3.3 percent and the second 1.7 km long and sloping at 3.8 percent occur (Figure 3). The reach downstream La Béarde is modelled due to active supply of material from the Tête de l'Aure slopes (2511 m), which, in turn, are modelled by debris flows. The lower reach of the Vénéon valley bottom corresponds with the valley broadening in Les Étages, being the terminal depression of the recessional phase M2 older than AD 1645. On the left slope of Les Étages, at heights 20 and 50 m above the channel, two lateral moraines are found whose age cannot be determined by lichenometry. The lower lateral moraine, developed in the form of a boulder rampart, is fossilized by an extensive fan formed at the outlet of the valley descending from Tête du Rouget summit. At present, the fan is inactive. The Vénéon channel in Les Étages dissects the bottom terrace to a depth of 5-8 m. It is a singular established channel, which does not show a lateral migration. From Les Étages, the Vénéon valley floor is inclined at 5.8 percent over a distance of 600 m. The next 800 m long reach, sloping at 3 percent is related to the terminal depression of Pièce du Clot. In the boundary zone, between the steeper (5.8 percent) and gentler (3 percent) reaches, there are two clear bends of the channel incised into bedrock. The width of the valley floor at Pièce du Clot is 150 m while the width between valley rocky walls is c. 500 m. In the valley bottom, the extensive level 7.5 m high above the present-day Vénéon channel bed and the ledge of the 5.5 m high terrace are distinguished. At the surface of the 7.5 m high terrace, covered with grass and singular trees, huge boulders occur. On these boulders the maximum measured lichen diameters are 110-100 mm (1570-1610). The blocks are likely to originate from rockfalls from the northern slope. The blocks could have fallen on the surface of the glacier filling up the terminal depression. At the level of the 5.5 m high terrace, lichen measured on singular large rounded pebbles are 45 mm in diameter (AD 1835), while at the upper scarp of the 7.5 m high terrace they are 40 mm in diameter (AD 1855) (Figure 11). The time of formation of the 5.5 m high alluvial level can be estimated at the years 1835 to 1855, and the dissection rate of this level at 33-38 mm/yr. Thus, it can be speculated that the dissection of the Pièce du Clot terminal depression by 2 m lasted until the beginning of the 19th century. Dissection of

valley bottom during 17-18th century seems to be very slow. The rate of incision grew after AD 1840.

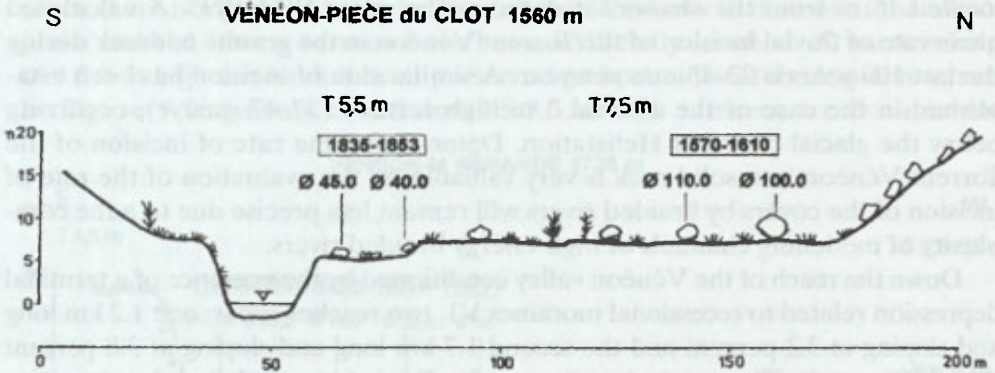


Figure 11. Cross-profiles of the Vénéon valley floor below confluence of the Étançons and upper Vénéon valleys.

In the longitudinal profile of the Vénéon valley (Figure 3) the maximum extent of the 14.8 km long glacier is marked. This limit is determined by an arc of the terminal moraine M1 preserved on the valley floor which forces the incised Vénéon channel into a distinct bend. The blocky moraine is on the left Vénéon bank. Unfortunately, measuring maximum lichen on this moraine was not successful but moraine seems to be older than AD 1570. Below moraine M1 the valley gradient is 1.3 percent.

FLUVIAL/GLACIFLUVIAL BRAID-PLAIN AND TERRACE LEVELS IN THE ÉTANÇONS VALLEY

The 8 km long Étançons valley is relatively broad when compared with the Vénéon valley and its width measured between the walls of glacial trough at their lower parts is about 250 m. In this valley, the upper 3.8 km long glacially modeled reach can be distinguished from the lower reach with an alluvial bottom and gradient of 7.3 percent. The alluvial reach can be subdivided into two parts: the higher one, 1.4 km long fed by a double, presently active fan of the Grande Ruine valley, and the lower one fed from both sides by active fans. The latter are associated with du Plaret and Aréna glaciers in the west and the hanging de la Sane valley, without a glacier at present. The Étançons stream pressed between these fans formed an extensive 200 m wide alluvial plain upstream and 150 m wide plain downstream. An inflection point marked in the Étançons longitudinal profile (Figure 4) indicated aggradation in the reach above the fans forced by the valley narrowing. The alluvial reach down valley from the narrowing is mainly fed by melting glaciers in the hanging valleys.

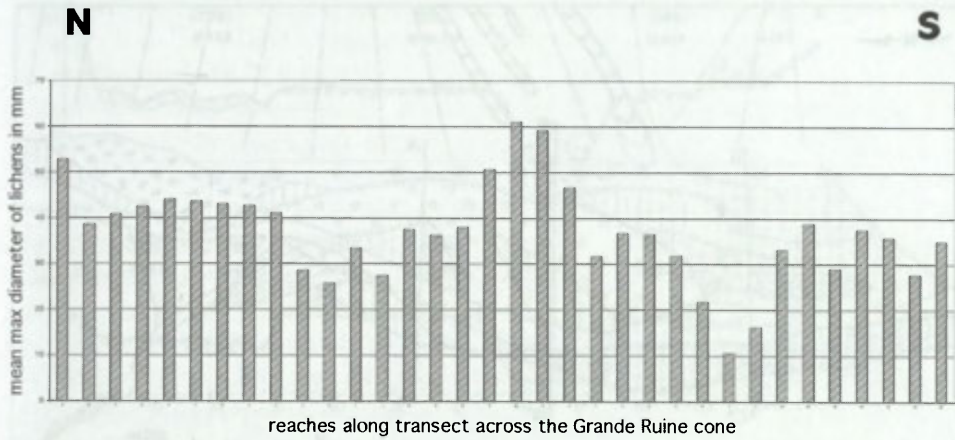


Figure 12. Lichenometric data showing distribution of *Rhizocarpon* lichens on N-S transect across the Grande Ruine cone.

A. The percentage of vari-aged material on both glacialfluvial fans formed at the foot of a rocky sill of the Grande Ruine glacier cirque has been examined. These fans formed due to intensive melting of the Grande Ruine glacier after the retreat of the Étançons glacier from this section. Because of non-uniform melting, conditioned by shading of the north-facing cirque walls, the Grande Ruine glacier formed two interlocking glacialfluvial fans, which pushed the Étançons stream onto the opposite western slopes in the valley reach from 2170 m to 2080 m a.s.l. The spatial distribution of the parts of the cones, which are uniform, and of the same age shows that they developed from AD 1770 (lichen diameters 61-59.1 mm) to 1805-1825 (lichen diameters 52.8-46.6 mm). The material of the years 1839-1851 (lichen diameters 44.0-40.8 mm) is preserved most commonly on the fans. After that period development of particular fans is more differentiated. Activity of the lower cone intensified after 1861 (lichen diameter 38.8 mm) and continued until 1902 (lichen diameter 27.7 mm). In this period the activity of the second cone was limited to its left side and lasted to AD 1912 (lichen diameter 25.7 mm). Due to the glacier retreat towards the north-facing slopes, the material was subsequently transported only in the lower gorge dissecting the cirque threshold of the Grande Ruine valley, and only the lower fan was developing. The youngest fragments on this fan are dated at 1962. In the central part of this fan, a trough with the 1 m high erosional level was incised. At present glacier waters use this trough. On the upper fan, fresh material is found on its right outermost fragments. The fan relief modelled in the periods mentioned above has been reconstructed based on the transects running in the lower parts of these fans, because in the middle parts of the fans and at their apices lichens of smaller diameters dominate which suggests accretion of these parts of the fans. During the development of the great fans of Grande Ruine, barricading the main valley,

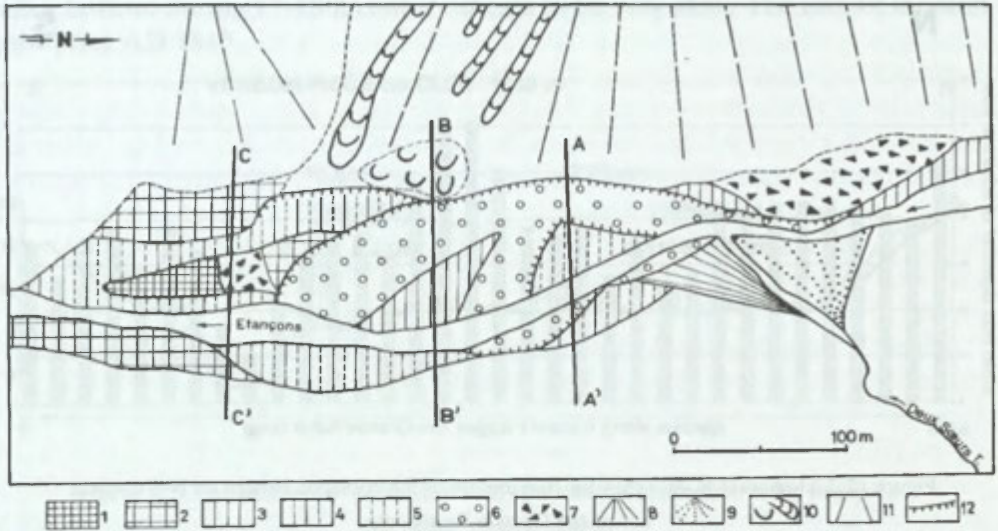


Figure 13. Valley floor landforms and floorplain age according to lichenometric data – Etançons braided system differentiated according to lichens sizes. Terrace levels: 1 – \varnothing 56 mm, 2 – \varnothing 42 mm, 3 – \varnothing 36-37 mm, 4 – \varnothing 28-30 mm, 5 – \varnothing 23 mm, 6 – \varnothing 13-15 mm, 7 – rockfall/rockslide deposits, 8 – older glacialfluvial cone – \varnothing 35 mm, 9 – younger glacialfluvial cone – 16 mm, 10 – debris flow levees and tongues, 11 – avalanche slope, 12 – erosion edges of terrace.

the head glacier of the Étançons valley was retreating from moraines E2 and E3, i.e. from the positions above and below Refuge du Châtelieret (alt. 2225 m) tourist shelter. Downvalley, from 2170 m to 2080 m, the valley bottom is relatively narrow over the 400 m long section, is 50-70 m wide, and dissects the level preserved at the foot of the right slopes shaped by avalanche activity (Figure 4).

B. Geomorphologic mapping has been carried out in the best developed section of the valley with the braided channels and the age of the forms determined by lichenometry (1960 – 1940 m a.s.l.). Figure 13 presents the setting of the channel, paleochannels and fragments of the alluvial level over the distance of 600 m and 150m width. Characteristics of the alluvial terrace levels, bars and channels are presented in three cross-sections (Figures 13 and 14). The channels are fed with alluvial material by the Étançons stream and by a long slope modelled by debris flows. In this reach the levels of the following heights can be distinguished: 3 m above the present-day channel, 2.5-2 m – which is most common, and 1.5-1 m. The latter levels have no shrub vegetation. The levels uncovered with vegetation are characterised by mean maximum *Rhizocarpon* lichen: 13.6, 14.5, 15.6, 18.4 mm. The higher levels partially covered with shrubs and bushes and trees can be characterised by lichen sizes: 37.0, 36.5, 29.4, 29.0, 28.3 but also of the size of 42.4 mm. That refers to 2.5-2 m high terraces. The 3 m high terrace fragment is overgrown with a compact vegetation cover and is adjacent to the fragment

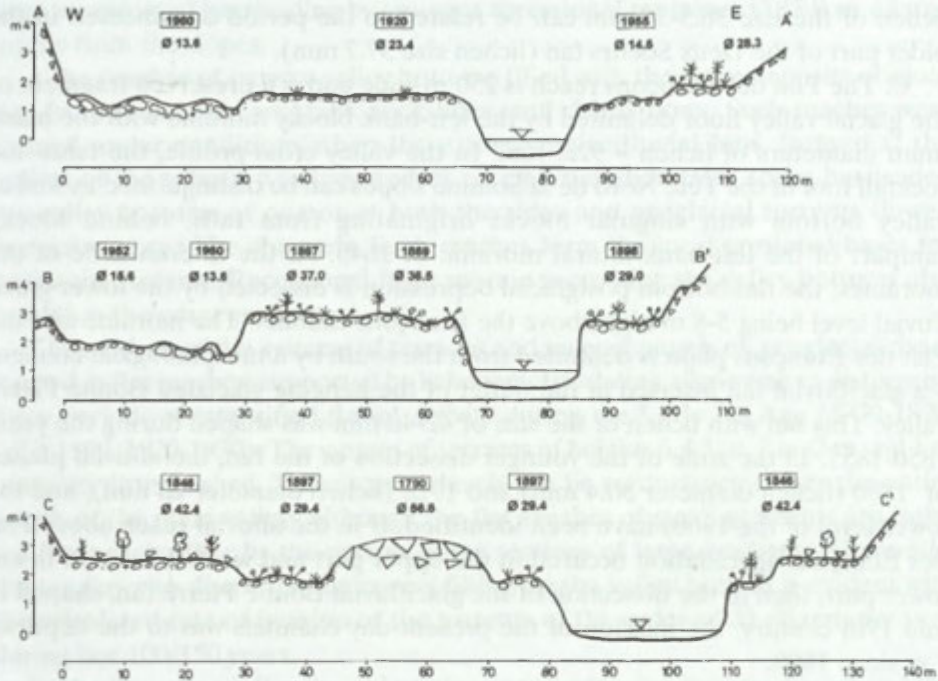


Figure 14. Cross-profiles of the Étançons valley floor located on figure 13.

of the boulder mound where lichen of the size of 56.6 mm have been measured. The sharp-edged blocks can be related either to a rockfall or to debris flow deposition. On the neighbouring slope there are levees of the debris flows with dead lichen 52.5 (lichen size 49-62 mm) (Figure 13). On the lower profile C the lichen size of 42.4 mm refers to 2.5 m high level. In the 15 m wide overgrown paleochannels, however, the lichen of the mean maximum of 29.4 mm have been measured. Thus, considering the 600 m long reach, one can state that in its upper part the younger fragments occur 1-2 m high above the present-day channel. These fragments were formed during the years 1955-1960, and a fragment of the 2 m high terrace originates from c. 1920. In profile B the paleochannels at the height of 2 m were modelled during the decade 1950-1960, while in profile C paleochannels at the height of 2 m were modelled earlier, i.e. about 1900. This indicates that in the upper part of the reach aggradation occurs while in the lower part, there is incision of the main channel when compared to the surface modelled in the period prior to 1850. This evidences also that in alimentation of the discussed alluvial reach longitudinal transportation predominates over the supply from the slopes.

The alluvia of the terrace formed c. 1850 (lichen diameter 40-42 mm) correspond with the most widespread reaches of the Grande Ruine fan dated at 1839-1851 (lichen diameter 44-40.8 mm). Fragments of the younger terrace with

lichen of the size 36.5-37 mm can be related to the period documented in the older part of the Deux Soeurs fan (lichen size 37.7 mm).

C. The Plat des Étançons reach is 250 m wide and is a preserved fragment of the glacial valley floor delimited by the left-bank blocky moraine with the maximum diameters of lichen – 92.5 mm. In the valley cross-profile, the talus and rockfall foot of the Tete Nord de la Somme slopes can be distinguished as well as valley bottom with singular blocks originating from falls, behind blocky rampart of the left-bank lateral moraine of 1645. At the internal side of the moraines, the flat-bottom postglacial depression is dissected by the lower glaci-fluvial level being 5-8 m high above the Étançons channel. The moraine and the Plat des Étançons plain is delimited from the south by a morphological element – a glaci-fluvial fan inserted at the outlet of the hanging glaciated Bonne Pierre valley. This fan with lichen of the size of 42-40 mm was shaped during the years 1850-1855. In the zone of the younger dissection of the fan, the alluvial phases of 1890 (lichen diameter 30.4 mm) and 1925 (lichen diameter 23 mm), and the lower level of the 1960s have been identified. If in the alluvial reach above Plat des Étançons aggradation occurred in the upper part and weak dissection in the lower part, then in the dissection of the glaci-fluvial Bonne Pierre fan, shaped in mid 19th century, the incision of the present-day channels was to the depth of 1 m since 1890.

CONCLUSIONS

The present-day relief of the alpine valley floors in the Écrins massif was formed as a result of interacting glacial, glaci-fluvial, fluvial and slope processes. The morphogenetic role of the several processes varied in time and space. Depending on the set of processes which predominated at the front of the retreating glaciers during the last 400 years and on the size of the alpine valleys, valley reaches were formed differing in genesis. An important role in the postglacial evolution of the relief of the main valley was the separation of the hanging tributary glaciers from the main ones by the cirque threshold and having cirque glaciers which are present until contemporary times.

Generally, two fundamental types of valley floor reaches can be distinguished in the extraglacial parts of the Vénéon and Étançons valleys;

- reaches of flat bottom valleys filled with the deposits of basal and ablation moraines, subjected to postglacial modelling by glaci-fluvial and fluvial waters. The bottoms of these reaches have a characteristic braided river pattern, subject to regular re-modelling during the floods associated with proglacial water and extreme hydrometeorological phenomena. Locally, lateral migration of the channels affected the whole valley floor with dissection and removal of the deposits filling these bottoms, giving a negative balance of the sediments. Aggradation occurs only in restricted places and points to a positive balance and arrises

from a preserved barricading by younger recessional moraines (1850) or a large supply from the slopes.

- the reaches of narrow valley bottoms filled with the slope deposits originating from substantial rockfalls, rockslides, and debris flows. Such reaches were formed under conditions when the extensive glacifluvial fans, formed at the outlets of the tributary valleys, and/or rockfall/rockslide talus cones barricaded the valley bottoms at one or at both the sides and proglacial torrents flowed in narrow gorge-like channels. Such reaches form the local erosional bases for proglacial streams. Recessional front moraines crossing the valley bottoms also condition the gorge reaches.

The analysis of the systems of terraces and paleochannels of proglacial rivers formed in flat reaches, supported by lichenometric dating allows one to distinguish three periods of intensified fluvial activity during the Little Ice Age (1770-1820, 1850-1890, 1920-1930). The system of terraces of heights 6-4.5 m, 3 m, 2 m and 1 m has been distinguished. These terrace levels can be reconstructed over the entire length of the main valleys although the flat reaches of small gradients are isolated from each other by the narrow gorge sections of large gradients. Generally, the tendency to dissecting the covers filling up the valley bottom is evident with the calculated rate of incision of the torrents of the order of 35-45 mm per year during last 100-150 years.

In the Étançons valley the oldest elements evidencing the glacifluvial and fluvial activity are preserved on the fans of the tributary valleys joining the main valley (Grande Ruine, Bonne Pierre and Deux Soeurs). The fans of Grande Ruine were formed from 1770-1780 (lichen diameter 61-59.1mm), when the main valley glacier stopped at the position of moraine E2. The subsequent phase of the glacier stagnation in Étançons (phase E3) and in Grande Ruine evidences the period 1805-1825 (lichen diameter 52.8-46.6 mm). In the Étançons valley the fragment of the 4-3 m high terrace is the oldest element. The age is calculated from lichen diameters measured on rockfall blocks or washed debris flow levees (lichen diameter 56.6 mm – 1790). Both rockfall blocks and debris flow levees are preserved at Plat des Étançons upward. Formation of c. 2.5 m high terrace plains with paleochannels took place during 1850, 1870 and 1895 (lichen diameters 42, 37, 30 mm). The large fragments of the terraces are preserved from the last period. The level of the 2 m high terrace, shaped during 1916-1920, is also important. Presented fragments of the terraces rise above the braided plain and the river ceased its activity during the decade 1950-1960 (lichen diameter 15-11 mm). After this period the present-day channel incised. It should be emphasised that the periods c. 1850 and 1870 contributed to formation of glacifluvial fan of Bonne Pierre while dissection of this large form started at c. 1890. The stages of dissection of this form can be related to 1890 and 1940.

In the Venéon valley the oldest terrace, 4.5-5 m high above the present-day channel, was formed during the years 1815-1862, has numerous paleochannels and is best developed near Refuge du Carrelet tourist shelter. During the period

of terrace formation the Pilatte glacier was at positions V4 and V5, i.e. in the distance of 1.6-2.3 km up the valley. The Carrelet plain represents the terrace levels associated with recessional stages of the moraines of 1815-1820 and with the recessional moraine of c. 1850. Dissection of the terrace took place during the years 1852-1895 (lichen diameters 40-30 mm) and was marked by formation of the plain with paleochannels at the height of 3 m above the present-day channel. The 2.5 m high terrace ledge, age of c. 1850, located in a near-slope position, has been preserved in the valley section called Pierre Chamoisseire. At that time terraces were also formed in the lower section of the valley (Pièce du Clot). Development of the terraces during 1875-1895 comprised accretion of the broad segment of the 3m high terrace in the Carrelet reach. In 1850 deepening of the valley from 2 to 3 m started and the braided mechanism of the valley floor development predominated. Downstream on the Vénéon river the role of the youngest downcutting in the modelling of the 1-2 m high terraces becomes more important (especially in Pierre Chamoisseire and La Bérarde reaches). The rates of the dissection of the terraces in the lower reach of the Vénéon (33 mm per year) and in La Bérarde (33-43 mm per year) are similar.

Development of the valley floors conditioned by the recession of glaciers in the main valleys points to a dominating role of climatic changes in formation and dissection of the terrace levels. The lateral alimentation during recession of main and cirque glaciers caused changes in sediment supply to the valley bottoms. Among the fans providing material to the bottom of the main valley large fans, 200-300 m high above the present-day channel can be distinguished. The extensive cirque glaciers fed these fans. As these glaciers diminished, the drainage directions in some hanging valleys changed. At present, the prevailing supply is from the hanging valleys, through the dissected cirque threshold as the lower part of glacial undercutting on the slopes of the main valleys. Younger fans are at a height of c. 100 m above the present-day bottoms of the main valleys.

In the majority of the broadened reaches, in which fragments of terraces and alluvial plains modelled by the braided rivers are preserved, output of the material predominated over the lateral supply. Aggradation is marked upstream, on the reaches barricaded by the system of large fans located opposite to each other and also in the upper reaches from which the glaciers retreated after 1850. In the case of large hanging tributary valleys, as the Étançons valley, a limited dissection of cirque thresholds does not promote output of the material from the hanging valley.

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INTRODUCTION

One of the most pronounced phenomena associated with the geomorphic response to environmental changes is the formation of permanent gullies, representing significant indicators of both landscape and climatic changes. Thanks to having the countries affected methodically by gully erosion in the past where the gully network may locally show backward-like features. The decrease density of gullies is linked with hill lands and lower uplands, representing a transitional step between plains or lower parts of hills on the one side and higher uplands

The study ...

The results ...

CONCLUSIONS

GULLY EVOLUTION IN THE MYJAVA HILL LAND IN THE SECOND HALF OF THE LAST MILLENNIUM IN THE CONTEXT OF THE CENTRAL-EUROPEAN AREA

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ABSTRACT: The author's investigations in the territory of the Myjava Hill Land, Slovakia revealed two periods of gullying in the course of the second half of the last millennium, the first dated to some time between the second half of the 16th century and the 1730s and the second approximately between the 1780s and the middle of the 19th century. Though the extensive forest clearance and expansion of farmland provided conditions favouring gullying, the triggering mechanism of the disastrous gully erosion were extreme rainfall and snowmelt events within the Little Ice Age (LIA). The comparison of gully formation phases identified in the study area with stages of gullying known from some other central-European countries suggests that gullying was not fully simultaneous across the region. The older phase identified in the Myjava Hill Land, does not have an equivalent in Germany, Poland, or Hungary and to a considerable degree in Czechia either.

KEY WORDS: gully evolution, periods of gullying, land use changes, climatic changes, Slovakia.

INTRODUCTION

One of the most profound phenomena associated with the geomorphic response to environmental changes is the formation of permanent gullies, representing significant indicators of both land use and climatic changes. Slovakia is among the countries affected markedly by gully erosion in the past where the gully network may locally show badland-like features. The densest network of gullies is linked with hill lands and lower uplands, representing a transitional step between plains or lower parts of basins on the one side and higher uplands

and mountains on the other. This intermediate, submountainous zone is built predominantly of less resistant sedimentary rocks, resulting in significant thickness of regolith, prone to erosion processes. The gradual settlement of these areas and their agricultural utilization also created conditions for the effective operation of gully erosion. Among territories the most affected by gully erosion in Slovakia is the Myjava Hill Land (Bučko and Mazúrová 1958), which is why this area was selected for the study of historical gully evolution.

The article has three objectives. The first is the reconstruction of the inception and subsequent development of the gully network in the territory of the Myjava Hill Land in the course of the second half of the last millennium, based on the identification of periods with increased activity of gully erosion and assessment of the geomorphic effects of gully erosion on the basis of detailed terrain investigation and interpretation of various historical sources. The second goal is to contribute to international discussion concerning the question of whether land use or climatic changes were more important for disastrous gully erosion (or the soil erosion as a whole). As is well known, both opinions have their followers and opponents. The primary role of land use change is advanced for example by Vogt (1953, 1958a,b), Richter and Sperling (1967), Hard (1976), Semmel (1961, 1995), Lang (2003), Kukulak (2003), whereas a leading role for heavy rainfalls and climatic changes is advocated for instance by Bork (1985, 1988, 1989), Bork et al. (1998) and Buraczyński (1989/1990). In relation to the influence of climate, the dominant emphasis is upon the question of the role of increased frequency of extreme events within the Little Ice Age (LIA) as it affects gully formation. The third aim is to contribute to the discussion concerning the simultaneous occurrence of gully erosion in the region of central Europe, claimed by Bork (1989), on the basis of comparison of the dating of periods of gully formation in the Myjava Hill Land with known phases of gully erosion in other central-European countries.

MATERIAL AND METHODS

The basic method used was to assess the causal connection of investigated phenomena, i.e., the cause-effect relation. The cause is represented by environmental changes, both land use and climatic, the effect is the geomorphic response of these changes. The article analyses this relation in both directions. The direction from cause to effect is linked with an assessment of the gain (or loss) of areas affected by gully erosion in connection with increase (or decrease) of the areal extent of farmland to the detriment (or in favour) of woodland. The direction from effect to cause arises with deciphering phases of catastrophic gully erosion in the course of colder and wetter climatic fluctuations on the basis of knowledge of geomorphic responses in the form of a network of relic, permanent gullies.

Source material was acquired by detailed field geomorphic investigation, by the analysis of old maps, aerial photographs and current topographic maps, as

well as by the study of various regional and local historical sources. Field investigation of selected areas was based on mapping at the scale of 1:10 000 and the complex documentation of historical geomorphic effect of gully erosion, namely of permanent gullies. Maps of the 1st (1782; 1:28 000), the 2nd (1837/1838; 1:28 000) and the 3rd (1882; 1:25 000) military mapping of the Old Hungarian territory within the Habsburg Monarchy, as well as the cadastral maps from the turn of the 19th and 20th centuries in the scale of 1:2 880, were analysed. Military maps indicated the spatial development of the land use pattern and gully networks over time and the relative age of gullies. Cadastral maps helped to confirm the relationships of gullies to former land use patterns. Analysis of aerial photographs and detailed topographic maps directed attention towards specific gully localities appropriate for the field research.

The historical literature (Varsik 1972, Horváth 1979) elucidated the settlement history and indirectly also the land use history; the original written accounts by local priests (e.g., Paulíny 1892, Bodnár 1926) helped indirectly with the relative dating of gully formation.

NATURAL CONDITIONS OF THE STUDY AREA

The Myjava Hill Land is situated in western Slovakia near the frontier with the Czech Republic. This area of 384 km² is a downfaulted zone between the flysch massif of the White Carpathians and the limestone-dolomite part of the horst of the Little Carpathians. Its character is mostly plateau-like with a relief of the order of 40-130 m. It comprises primarily flysch-like rocks of medium to low resistance with a considerable thickness of fine-textured regolith. Patches of loess loams of variable extent and thickness, usually situated on lower slope portions, suggest that a more extensive loess cover existed in this area in the past. Loamy to clayey-loamy deluvium is often 10-15 m thick, exceptionally even more. The thickest beds of this material are situated on the foothills and in the bottoms of dells and dry valleys. Cambisols and Luvisols are the most frequent soil types. Mean annual precipitation is 650-700 mm, and the natural vegetation was predominantly oak and oak-hornbeam forest with beech forest in the highest parts (Stankoviánsky 1997a).

ENVIRONMENTAL CHANGES CONTROLLING GULLY FORMATION IN THE STUDY AREA

LAND USE CHANGES

The majority of the Myjava Hill Land was still forested and unsettled in the middle of the 16th century (Figure 1). The older settlements were concentrated almost exclusively in the marginal areas. From the first historical references the

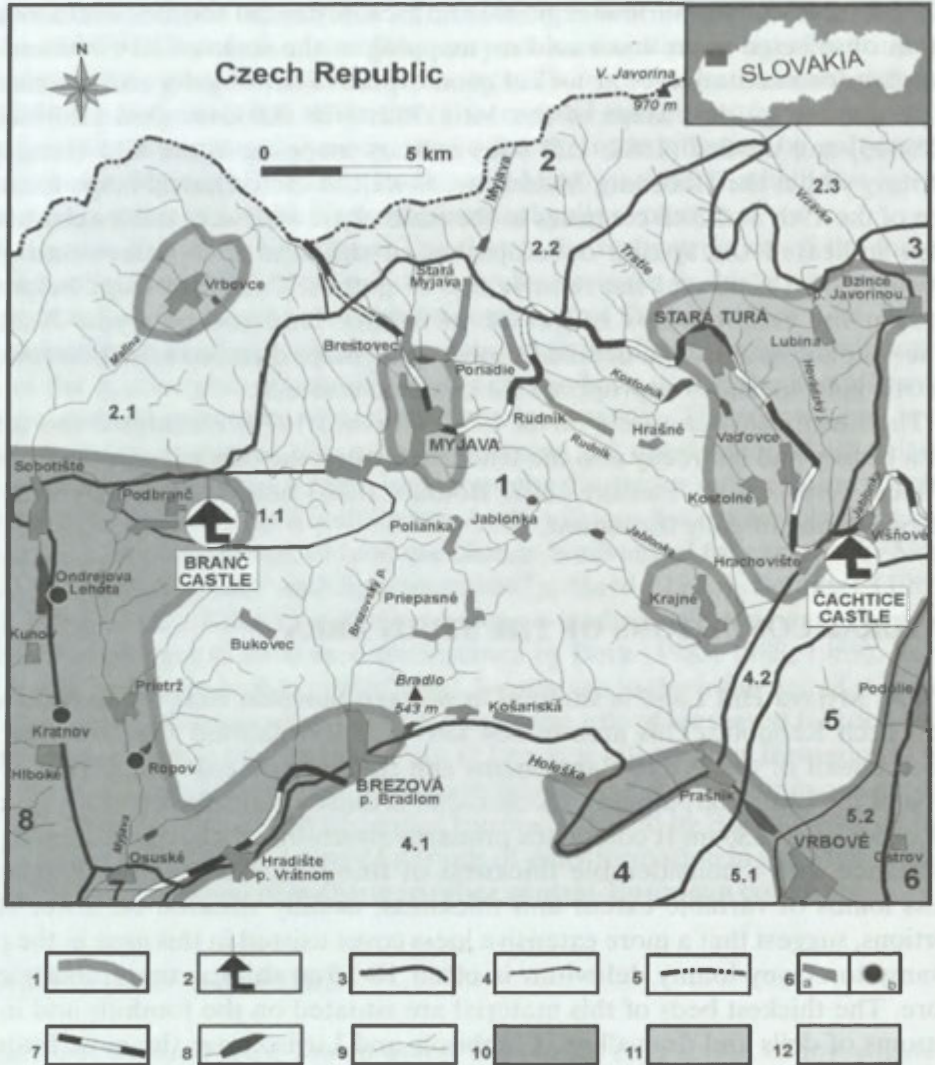


Figure 1. Map of settlement history in the Myjava Hill Land.

Legend: 1. boundaries of areas settled in different periods, 2. medieval castles, 3. boundaries of geomorphic units, 4. boundaries of geomorphic subunits, 5. state boundary with the Czech Republic, 6. a) present-day settlements, b) villages abandoned in the 14th and 15th centuries and not settled later, 7. railway roads, 8. water reservoirs, 9. streams, 10. territory settled in the 13th century and earlier, 11. territory settled in the 14th century, 12. territory settled in the period from the 2nd half of the 16th century to the end of the 19th century (the kopanitse colonization).

Geomorphic units: 1. Myjava Hill Land: 1.1. Branč Klippes; 2. White Carpathians: 2.1.

Zalostina Upland, 2.2. Javorina Highland, 2.3. Bosaca Klippes; 3. Povazie Basin; 4. Little Carpathians: 4.1. Brezová Carpathians, 4.2. Cachtice Carpathians; 5. Trnava Hill Land: 5.1. Sub-Little-Carpathian Hill Land, 5.2. Trnava Loess Plain; 6. Lower Vah Flood Plain; 7. Bor Lowland; 8. Chvojnica Hill Land.

settlement of the western and southwestern margin was known as early as the middle of the 13th century, but archaeological survey revealed that its beginnings were in the Slavonic period or even in the Neolithic era. Much later settlement occurred in the eastern parts of the study area, associated with the construction of the medieval Čachtice Castle (1263-1273). Numerous villages were founded there sometime between the second half of the 13th century and the first half of the 14th century. A little later, approximately at the turn of the 14th and 15th centuries, two large farms (praedia) were established near the present-day town of Myjava.

The principal stage of settlement was represented by the so-called 'kopanitse' colonization that started in the second half of the 16th century and culminated at the turn of the 18th and 19th centuries (Figure 1). During this period, lasting approximately 250 years, forest clearance of most of the settled territory and the subsequent transformation of the acquired areas into farmland took place. Deforestation was carried out, as during older colonization phases, by slash and burn technology and by hoe adjustment of the acquired ground for sustained farming or settlement purposes (Podolak 1966). The areal extent of the territory colonized, deforested and agriculturally used by the kopanitse settlers is approximately four times larger than the area colonized and cleared after the construction of the Čachtice Castle. Only some islands of original widespread forests were preserved. Deforested territory reached its maximum extent at the time of the culmination of the kopanitse settlement. During the 19th century, especially in its second half, afforestation started to increase gradually (although its origins reach locally back to the first half of the 17th century according to Mazúr 1970), and continued in the 20th century.

In connection with the limits of further areal increase of farmland at the expense of woodland after the culmination of the kopanitse colonization, and with the ever growing number of inhabitants, arable land gradually replaced widespread pastures and meadows and the hereditary division of the existing fields into ever smaller plots occurred. Thus, the characteristic feature of the kopanitse landscape in the Myjava Hill Land in the course of the 19th century, under the influence of the enduring farm fragmentation, was a mosaic of small, narrow fields, tilled predominantly along the contour lines, less often down-gradient, separated by the artificial linear landscape elements, as for example balks, lynchets, access roads or paths. This landscape type persisted until the middle of the 20th century, when the transformation into the contemporary farmland, characterised by large blocks of cooperative fields took place in connection with the collectivization in agriculture.

CLIMATE CHANGES

While we have a relatively good picture of the land use changes of the Myjava Hill Land from the beginnings of the settlement until present, our knowledge of climate changes over this period is very poor. This is due to the fact that the

reconstruction of climate in Slovakia is very recent (Brazdil and Kiss 2001). We could use only the conclusions of historical climatologists dealing with the evolution of climate in Europe as a whole or in its particular regions, with special regard to the territory of the Czech Republic, and the results of Czech colleagues helped us to produce at least an approximate picture of the chronology of climate fluctuations for the study area, as it is situated close to Czechia.

In the Myjava Hill Land, climate changes could start to manifest themselves geomorphically (by means of changes in frequency of extreme erosion events) only after deforestation. As the beginning of more extensive forest clearance associated with the kopanitse colonization is not older than from the second half of the 16th century, we are in fact concerned with the climate evolution which took place in the second half of the last millennium. According to Starkel (2000), of all climatic fluctuations, it is the wetter and at the same time colder periods typical of increased concentration of extreme meteorological-hydrological events, both rainfall and snowmelt, that exhibit a significant geomorphic effect. The same author (Starkel 2000, 2002) distinguishes four types of such extreme events. The first is represented by local heavy downpours of short duration, when the total rainfall involved may range from as little as 20-50 to 100 mm and more, but their intensity exceeds 1-3 mm per minute. The second type, affecting extensive areas of many thousands of km², is represented by continuous rainfall. In spite of its low intensity this type, due to its long duration, has high rainfall totals, reaching up to 300-1 000 mm within 2-5 days. The third type comprises rainy seasons when for several months or even a whole year the groundwater storage is close to 100 percent. The fourth type is represented by rapid snowmelt. Geomorphic effects of extreme situations increase where two events of different type are superimposed, above all when continuous rain is concluded by a high-intensity downpour (Starkel 2002) or when snowmelt is accelerated by simultaneous rain (Starkel 1976). Thus it is obvious that extreme events need not all show an equally 'extreme' geomorphic effectivity, because they differ in their extremeness according to the intensity of the particular type manifestations.

Of the known climatic periods of the last millennium (Brazdil 2000), landscape evolution in the greater part of the Myjava Hill Land, settled in the course of the kopanitse colonization, was under the influence of the LIA and that of the subsequent period of global warming. However, the most profound impact was left by the LIA, taking place according to Lamb (1977) in the period 1550-1850 or according to Flohn (1982) in the period 1570-1860. Though later discussion located the beginning of the LIA back to the period shortly after 1300 (Pfister et al. 1996, 1998) or even back to 1250 (Porter 1986), there has not been similar debate about identification of its end; according to different authors its dating ranges between 1850 and 1890 (Brázdil 1996). Thus, the part of the study area, colonized by the kopanitse settlers, experienced the LIA in the period from the second half of the 16th until the second half of the 19th century.

Similar shift of opinions also applies to the question of its internal division. Investigations confirmed unequivocally that at the lower hierarchical level the LIA divides into colder and wetter fluctuations separated by warmer periods (Jones and Bradley 1992). Moreover, today hardly anybody still doubts the simultaneity of oscillations within this very specific era of colder climate, not only from the global but also continental (in our case European) viewpoint. Diversity of opinions on this question is a manifestation of considerable spatio-temporal variability in climatic anomalies within the LIA, referred to numerous authors, e.g., Jones and Bradley 1992, Brazdil 2000, Starkel 2000. This is why we can only reiterate the statement of the last mentioned author that 'the phase of the LIA is well expressed throughout the Europe, although the particular events were never simultaneous'. What is for us in this sense the most important, undeniable fact, is that the territory of the Myjava Hill Land experienced the LIA and that the associated concentrations of extreme rainfall and/or snowmelt events, manifested themselves by significant geomorphic effects.

RESULTS AND DISCUSSION

INFLUENCE OF LAND USE CHANGES AND EXTREME EVENTS ON SPATIO-TEMPORAL DEVELOPMENT OF GULLY NETWORKS IN THE STUDY AREA

Detailed field geomorphic investigations revealed that the majority of gullies in the study area were conditioned by the economic activity of man; these gullies could be formed only during or after the transformation of woodland into farmland (Stankoviansky 2000). The gullies literally follow the farmland pattern, mostly the artificial linear landscape elements, from the time of their origin (Figure 2).

Most gullies are well preserved, sharply defined with a V-shaped cross profile (Figure 3). Gully depth is commonly from some meters up to 10-15 m, less frequently up to 20 m and exceptionally even a little more. Gullies are either isolated or in sets. The gully sets occurring on straight slopes are arranged in a typical parallel pattern (Figure 4), but those in dry valley heads are arranged in a dendritic pattern. The areas with the densest concentrations of gullies (up to 11 km per km²) exhibit badland-like features (Stankoviansky 1997b).

Knowledge of the settlement and land use evolution provided the opportunity to determine the lower temporal boundary for gully formation. The relatively late settlement for most of the study area, namely during the kopanitse colonisation, represents a great advantage for researchers as it unequivocally confirms that gullies in this area could not be formed earlier than in the second half of the 16th century. On the basis of the analysis of old maps and the local historical sources two periods of possible gully formation in the area colonised by the kopanitse settlers were identified, the first dates to some time between the end of the 16th century and the 1730s and second approximately between the 1780s and the



Figure 2. Geomorphic map of the dry valley catchment westwards of the village of Kostolné (including the relic, permanent gully network controlled by artificial linear elements of the original landscape).

- Legend: 1. flat bottom of main valley, 2. dry valleys, 3. remnants of older bottoms in dry valleys, 4. washed furrows incised in older bottoms of dry valleys, 5. shallow linear slope hollows of various origin, 6. permanent gullies, 7. colluvial fans, 8. landslides, 9. remnants of planation surfaces, 10. structurally conditioned elevations, 11. steps of old terraced fields, 12. stony agrarian elevations (a. in the form of terraced steps, b. in the form of ridges)

1840s inclusive (Stankoviansky 2003a). There are some indications that those parts of the study area, which were settled earlier could be affected by gully formation even before the older of the identified periods (Stankoviansky 2003b), but the assessment of still older phases of gullying was not among the goals of this article.



Figure 3. Well preserved permanent gully stabilised by afforestation

Gullies in the study area were formed mostly by vertical incision of ephemeral flows concentrating during extreme rainfall and sudden snowmelt events along the axes of the linear landscape features, both topographical and artificial, with the latter playing a much more important role. Thus, the factor responsible for the gully formation was concentrated runoff from rainfall or snowmelt waters.

In general, gullies are formed in a relatively short time (Moldenhauer 1995), obviously during some consecutive extreme events. According to Zachar (1970, pp. 356-357), they can be formed during 1-2 downpours, according to Bork (1989) and Stehlík (1954) exclusively even during a single event. What was the gully morphogenesis in the study area in the light of their supposed initiation and development?

As the forest clearance in the territories colonized by the kopanitse settlers was a progressive and long-term process, the pioneer gullies, incised during the first events of disastrous gullying could be created only in those limited areas that were already deforested. This scheme was valid especially for the first period of gully formation lasting from the second half of the 16th century until the 1730s. Thus, during later extreme events, both concentrated and isolated, new gullies could be formed mostly on the newly acquired farmland, while for the earlier deforested areas further growth of existing gullies was more common. In general, the type and magnitude of the particular event influenced the extent of gully erosion. Gullies formed in this period are recorded on maps from 1782. A different situation was that of the second of the identified periods of possible gully formation, lasting from the 1780s until the middle of the 19th century.



Figure 4. Aerial photograph from 1955 illustrating the land use pattern before collectivization. Forest in top half masks the gully network of the highest density in terms of the study area (11 km per km²), the right bottom quadrant shows the parallel system of gullies formed on former pasture.

As the forest clearance reached its maximum at the turn of the 18th and 19th centuries, new gullies could be formed almost exclusively in areas deforested for the longest time. Gullies from this period are recorded on maps from 1837/1838 or 1882. However, the latter maps depict predominantly only growth of existing gullies and not the formation of new ones.

Thus, the research results show that the overwhelming majority of gullies (especially the bigger ones) in the study area were formed in stages. This conclusion is in accordance with opinion of Zachar (1970, p. 357) as regards the general predominance of stage-like gully formation. Growth of gullies by headward erosion followed during each extreme event, or at least during the most significant of them. Some gullies approached divides gradually, in exceptional cases even penetrated them.

Current knowledge does not enable the absolute dating of individual occurrences of extreme events, both in clusters and single, leading to the formation of pioneer gullies and their further growth in the study area. However, we have come to an understanding of regularities of spatial evolution of gully network, or

in other words to a certain scheme, characteristic for the three centuries lasting era of irregularly repeated occurrence of disastrous gullying. We identified two periods of possible gully formation in this era on the basis of the analysis of historical sources. It is obvious that extreme meteorological-hydrological events, enabling the gully formation, had to occur some times during both of these periods.

ROLE OF CLIMATIC FLUCTUATIONS WITHIN THE LIA IN GULLY FORMATION IN THE STUDY AREA

According to Morgan (1995, p. 20), the main cause of gully formation is in general too much water, a condition which may be brought about by either climate change or alteration in land use. Bučko and Mazúrová (1958) and Klimaszewski (1981, p. 297) suggested that the genesis and evolution of gullies in the forested midlatitude morphoclimatic zone of Europe was associated especially with extreme rainfall and sudden snowmelt, though the conditions finally leading to gully formation in this zone were usually deforestation and agricultural utilisation of a land with deep soft regolith. All these conditions were met in the Myjava Hill Land. Intervention of man in the landscape in the course of the settlement, above all during the kopanitse colonisation, exposed the area to the influence of extreme events. The period of the kopanitse colonisation with accompanying land use changes overlaps in fact temporally with the period of the LIA (Stankoviansky 1997b). The fact that the overwhelming majority of larger gullies in the study area were formed between the middle of the 16th and the middle of the 19th centuries suggests that the decisive role in gully formation was played by the climatic factor. In other words, extensive forest clearance by kopanitse settlers and expansion of the farmland could have predisposed the land to gullying, but the triggering mechanism of the disastrous gully erosion may have been represented by extreme events during the LIA.

The LIA was not continuously cold and wet, but colder and wetter fluctuations occurred in certain time intervals. To look for the coincidence of the phases of gullying and climatic changes, let us try to relate to periods of possible gully formation identified in the Myjava Hill Land, with climatic fluctuations within the LIA identified by historical climatologists and others.

The older of periods of probable genesis of gullies in the study area from the second half of the 16th century to the 1730s overlaps almost completely with the most pronounced phase of the LIA in the years 1550-1700 identified by Lamb (1977). Brázdil (1996) identified a more humid phase with greater annual precipitation totals in the same interval in neighbouring Czechia on the basis of assessment of running averages. However, this phase was not continuously cold and wet. According to Starkel (2000), the wetter fluctuation with the greatest spatial extent in European conditions occurred in the years 1560-1570. Pfister and Brázdil (1999) identified the colder and wetter period in central Europe in the years 1585-1597, Brázdil and Kotyza (2001) in the years 1568-1599. According to Brázdil (1992), the extremely cold and humid period in Czechia occurred in the

last decade of the 16th century. This interval correlates well with the coldest of fluctuations identified by Briffa and Schweingruber (1992) on the basis of dendroclimatological analyses in the years 1590-1610 (other colder intervals identified by these authors, were in the years 1670-1680 and 1705-1720). According to Starkel (2000), cold fluctuations are periods with high concentration of extreme rainfalls. Glaser et al. (2000), relying on annual precipitation averages identify wetter climatic oscillation in the years 1640-1740, with a maximum in 1650. The results of dendroclimatological analyses by Brázdil et al. (2002) from southern Moravia, confirm wetter fluctuations in the neighbouring territory to the Myjava Hill Land in decades 1670-1680 and 1710-1720. These more humid decades apparently coincide with above cold fluctuations reported by Briffa and Schweingruber (1992).

These facts suggest that at least some colder and above all more humid fluctuations occurred in Europe and most importantly even in the neighbouring Czech Republic in the course of the older of the two identified periods of possible gully formation in the Myjava Hill Land. It is obvious that some of these fluctuations (or maybe also any of so far unknown fluctuations) resulted in the disastrous gullying in the study area. More precise and more unequivocal assessment will be possible only after the reconstruction of climatic evolution in Slovakia and particularly in the study area.

The younger period of possible gully formation in the Myjava Hill Land between the 1780s and the middle of the 19th century evidently coincides with the last phase of the LIA, which according to Lamb (1984) and Brázdil (1996) ended in the mid 19th century. The beginning of this phase, according to Lamb (1977), resulted from the revival of catastrophic rainstorms at the turn of the 18th and 19th centuries. Also during this period some partial colder and wetter fluctuations were identified. According to Briffa and Schweingruber (1992), such fluctuation occurred in the years 1800-1805, according to Brázdil (1994), in Czechia in the years 1780-1810 and 1820-1860. The latter author also identified two colder fluctuations between the older and younger periods of possible gullying in the study area, namely in the years 1730-1740 and 1750-1770. However, according to Brázdil et al. (2002), just the 18th century was the driest and the 19th century was the wettest in southern Moravia (close to the Myjava Hill Land) in the last six centuries.

GULLY EVOLUTION IN THE STUDY AREA IN THE CONTEXT OF PHASES OF GULLYING IDENTIFIED IN CENTRAL EUROPE

The two periods of possible gully formation identified in the Myjava Hill Land can be compared with results from another central-European countries, namely Germany, Poland, Czech Republic and Hungary. According to Bork (1989), gullies shown on the present-day German large-scale maps developed mainly during the second half of the 18th century and this period represented the only phase of significant gully erosion within the second half of the last mil-

lennium in almost all parts of central Europe. However, he later admitted that some gullies were also formed in the late 17th century, early 18th and early 19th centuries (Bork et al. 1998, p. 261). Hard (1976) estimated that gullies in SW Germany were formed between 1750 and 1850, and Buraczyński (1989/1990) regarded the present gullies in Goraj Roztocze (Poland) as the result of a phase of gullying starting at the turn of the 18th and 19th centuries. The same age was ascribed to gullies in the marginal zone of the Carpathian Foothills near Kraków (Poland) by Pietrzak (2000). In Czechia, Stehlík (1981) identified the phase of accelerated erosion like Hard in Germany in the 1750-1850 period. This coincides with the conclusions of Laznicka (1957), who on the basis of the analysis of old maps documented the growth of existing gullies in the Jihlava River vale (southern Moravia, Czech Republic) in the period 1785-1877. However, contrary to Stehlík, Zachar (1970, p. 332) on the basis of the study of historical sources found that in the Rakovník region (western Czechia), the majority of larger local gullies were formed in the 17th century and only a minority date to the 18th century. An interesting situation was found out on the basis of the analysis of old maps in the hilly territory of the Rakaca Catchment (58 km²) in Hungary. This area was gully-free before 1784. In the period 1784-1860 gullies in a total length of 21 km were formed, while in the period 1860-1920 the quoted length almost doubled (38 km) and the most intensive growth of gullies occurred in the period 1920-1994, when their total length reached 71 km (Gábris et al. 2003).

This compilation suggests that the phases of gullying in the second half of the last Millennium in the central Europe were not fully simultaneous across the region, in contradiction with the generally accepted opinion of Bork (1989). The older of the two phases identified in the study area, has no equivalent either in Germany (Hard 1976, Bork 1989), in Poland (Buraczyński 1989/1990, Pietrzak 2000), or in Hungary (Gábris et al. 2003), and if we strictly adhere to Stehlík (1981) not in the Czech Republic.

Bork (1989), Buraczyński (1989/1990) and Pietrzak (2000) attributed gully formation to climatic fluctuations during the LIA. Our results also confirm the influence of these fluctuations on gully formation. However, the stated regional differences in dating gullying suggest that the phases of catastrophic gully erosion in areas of Germany, Poland, Czech and Slovak Republics and Hungary need not be and obviously were not simultaneous, as the climatic fluctuations, which conditioned gully erosion, were not simultaneous either.

Although we attribute two phases of gully formation to climatic fluctuations, we do not exclude gully origin either in the intervening period, or more recently. According to Maruszczak (1973), gully development is sporadic, with periods of minor activity every several years during spring snowmelts and major periods every several decades during periods of increased rainfall. In the South Polish uplands, for instance, disastrous rainfalls occur 2-3 times a century (Maruszczak 1986). Stehlík (1954) introduced an illustrative example of such an event in the environs of the town of Bzenec in south Moravia (Czech Republic). The early

morning downpour on June 9, 1953 lasting two hours resulted, besides other damages, not only in the formation of 1-2 m deep erosion furrows controlled by balks between gradient fields, but also up to 7-8 m deep gullies linked to access roads and cart-tracks. It was the effect of the single event, though this particular erosion was encouraged by the fact the ground was waterlogged by numerous downpours on preceding days. Extreme events not related to the LIA were also responsible for the above-mentioned gully formation in NE Hungary after 1860 and especially for its acceleration after 1920.

Though extreme rainfall events leading to gullying can occur not only during climatic fluctuations, it is obvious that the long-term period with increased precipitation totals and greater probability of increased frequency of significant events provide more opportunities for gully formation. According to Starkel (1998, 2000), the ideal conditions for rapid development of gullies are especially represented by repetition of the sets of extreme rains, which were frequent in the course of the LIA. This standpoint is also supported by our investigation based on comparison of gully networks on the maps of the 2nd and 3rd military mapping with geomorphic maps reflecting current gully patterns. This comparison suggests a marked reduction of geomorphic effects of gully erosion after the end of LIA, as few new gullies originated after that time. Only a very limited local growth of the existing gullies took place, which could be conditioned by isolated events with limited reach in terms of area. Though the marked drop in growth of gully network was to some measure also influenced by afforestation, we attribute the key cause of this phenomenon to climatic change, at the end of the LIA.

CONCLUSIONS

Detailed field investigation, analysis of old maps and interpretation of written historical sources made it possible to identify two periods of gullying in the territory of the Myjava Hill Land in the course of the kopanitse colonisation (taking place from the second half of the 16th until the end of the 18th century) and after it, the first dated to some time between the second half of the 16th century and the 1730s and the second approximately between the 1780s and the middle of the 19th century.

The period of the kopanitse colonisation with the accompanying land use changes broadly corresponds with the LIA. The fact that the overwhelming majority of larger gullies were created during the above-mentioned two phases of gully erosion in the LIA suggests that the decisive role in their formation was played by the climatic factor. Thus, the extensive forest clearance by the kopanitse settlers and expansion of farmland provided conditions favouring gullying, but the triggering mechanisms of the substantial gully erosion were the extreme rainfall and snowmelt events, which occurred according to Starkel (1998, 2000) more frequently during colder and wetter fluctuations within the LIA. These

conclusions supplement other work pointing to the role of extreme events during the LIA in accelerating various geomorphic processes in central Europe, as for example soil erosion (Starkel 1994), gravitational (Kotarba 1995) or fluvial processes (Pišút 2002).

The comparison of gully formation phases during the LIA identified in the Myjava Hill Land with the stages of gullying known from some other central-European countries for the same period suggests that gullying was not simultaneous across the region. The older phases identified in the study area, do not have an equivalent in Germany, Poland, Hungary and to a considerable degree in the Czech Republic. These conclusions contrast with the generally accepted statement of Bork (1989) on synchronism of gully formation in central Europe but confirm the standpoint of Starkel (2000) that manifestations of the LIA were not simultaneous.

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ABSTRACT: Human activities have affected the paleoenvironmental system in the South German low relief hills and adjacent areas over centuries. The Holocene landscape history, influenced by human activities, cannot only be reconstructed if colluvial deposits, soils and landscape evolution (erosion, accumulation) of such and large river valleys are investigated in a complex way: the traces of the reorganization of colluvial deposits and their landscape as dominant system surface from the formation of the floodplain remnants. The first, being behind, accumulation on the hills and in the floodplain areas depends on the magnitude and morphology of the sediment. A reworked soil and colluvial system illustrates the different parts of the sediment. An important system is the colluvial deposits, however, the beginning of the landscape evolution through, but they cannot record climate changes. On the other hand, landscape evolution can be used to evaluate the age, origin and history of landscape erosion.

KEY WORDS: colluvial deposits, Holocene landscape, erosion, accumulation, erosion, erosion system, paleoenvironmental system, geobotany, soil erosion, landscape evolution.

INTRODUCTION

The cause of the changes of ecosystems, especially linear systems, and their extent, have been the subject of critical discussion by Lotze Starker for several decades (Starker 1960, 1999, 2002a,b). Holocene changes of ecosystems as a function of societies (settlement processes) and climate play a critical role in

HUMAN AND CLIMATE IMPACTS ON THE HOLOCENE LANDSCAPE DEVELOPMENT IN SOUTHERN GERMANY

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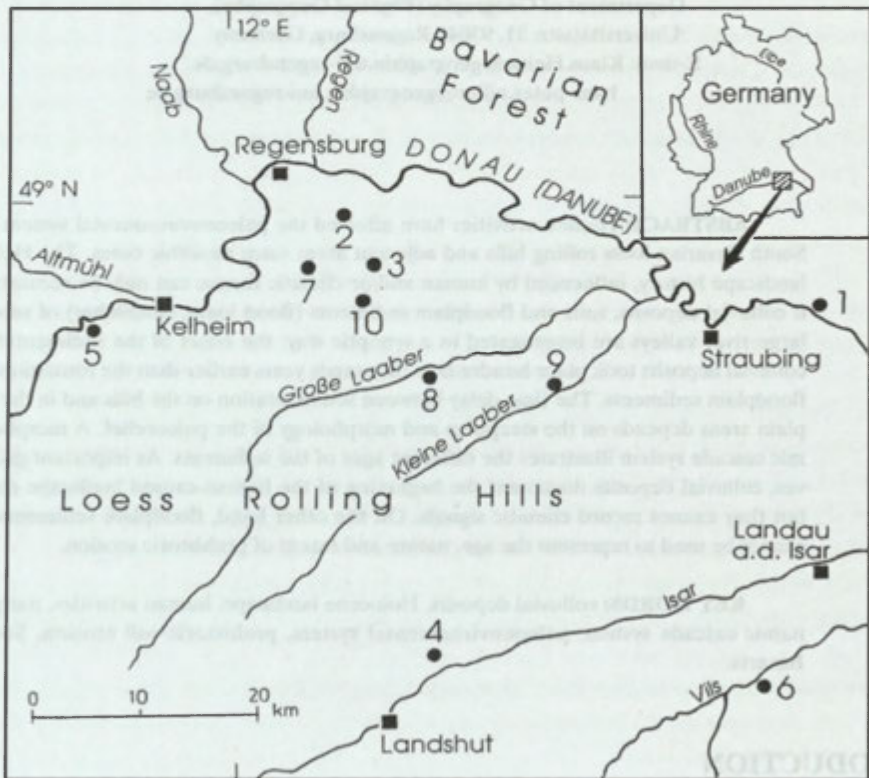
ABSTRACT: Human activities have affected the paleoenvironmental system in the South Bavarian loess rolling hills and adjacent areas since neolithic times. The Holocene landscape history, influenced by human and/or climatic forces, can only be reconstructed if colluvial deposits, soils and floodplain sediments (flood loam, Auenlehm) of small and large river valleys are investigated in a synoptic way: the onset of the sedimentation of colluvial deposits took place hundreds to thousands years earlier than the formation of the floodplain sediments. The time delay between sedimentation on the hills and in the floodplain areas depends on the steepness and morphology of the paleorelief. A morphodynamic cascade system illustrates the different ages of the sediments. As important geoarchives, colluvial deposits document the beginning of the human-caused landscape changes, but they cannot record climatic signals. On the other hand, floodplain sediments alone cannot be used to represent the age, nature and extent of prehistoric erosion.

KEY WORDS: colluvial deposits, Holocene landscape, human activities, morphodynamic cascade system, paleoenvironmental system, prehistoric soil erosion, Southern Bavaria.

INTRODUCTION

The cause of the changes of ecosystems, especially fluvial systems, and their extent, have been the subject of critical discussion by Leszek Starkel for several decades (Starkel 1960, 1999, 2002a,b). Holocene changes of ecosystems as a function of societies (settlement processes) and climate play a critical role in

Global Change research. Human changes to the earth system do not operate in simple cause-effect relationships. Investigation of the complex interactions between man and environment needs a multivariate and multidisciplinary approach (Leopold 2003). To reconstruct past environmental conditions it is important that several different methods are used to provide a collaborative approach both to validating methods and in determining the driving forces of geomorphodynamic changes. The South Bavarian loess area, the Danube valley with its tributary valleys and the southern slopes of the western Bavarian Forest (Figure 1) have been densely populated since prehistoric times. Thus, this area can be used as a landscape model to explore the consequences of population dynamics and/or climatic fluctuations for landscape evolution. Phases of denudation and accelerated soil erosion, of floodplain sedimentation and of soil development illustrate how humans and climate have contributed to environmental changes since late glacial times.



- | | | |
|----------------|--------------|------------------------|
| 1 Bogenberg | 5 Frauenberg | 8 Große Laaber valley |
| 2 Burgweinting | 6 Reisbach | 9 Kleine Laaber valley |
| 3 Eggfing | 7 Seedorf | 10 Thalmassing |
| 4 Essenbach | | |

Figure 1. Map of the study area with locations.

One documented result of the role of land-use change by humans is accelerated soil erosion on fields and sedimentation of the eroded soil material in floodplain areas (Rathjens 1978, Bork et al. 1998, Küster 1995). Data is presented here that supports the conclusion that, since neolithic times, human activities have affected the environmental systems, especially the soil erosion and accumulation processes, in an extent that was not previously known from the study area.

In this study, geomorphologic, sedimentologic and pedologic research on colluvial deposits and soils together with prehistoric and historic investigations should elucidate the following questions:

- Can a high time resolution of environmental changes for certain Holocene phases (e.g., c. 3,500-1,500 yr cal. BP) be achieved by using colluvial sediments and soils as paleoenvironmental archives?
- Are paleoclimate signals documented in colluvial sediments and soils?
- How and to what extent did human land-use and settlements influence soil erosion and deposition processes as well as soil formation?
- Additionally, our research has revealed that the landscape history can only be reconstructed if colluvial sediments, soils and floodplain sediments of small and large valleys are considered together. These and other items were the subject of a programme of the German Science Foundation (DFG Priority Programme 'Changes of the Geo-Biosphere during the last 15,000 Years. Continental Sediments as Evidence for Changing Environmental Conditions') (Zolitschka et al. 2003).

STUDY AREA AND METHODOLOGY

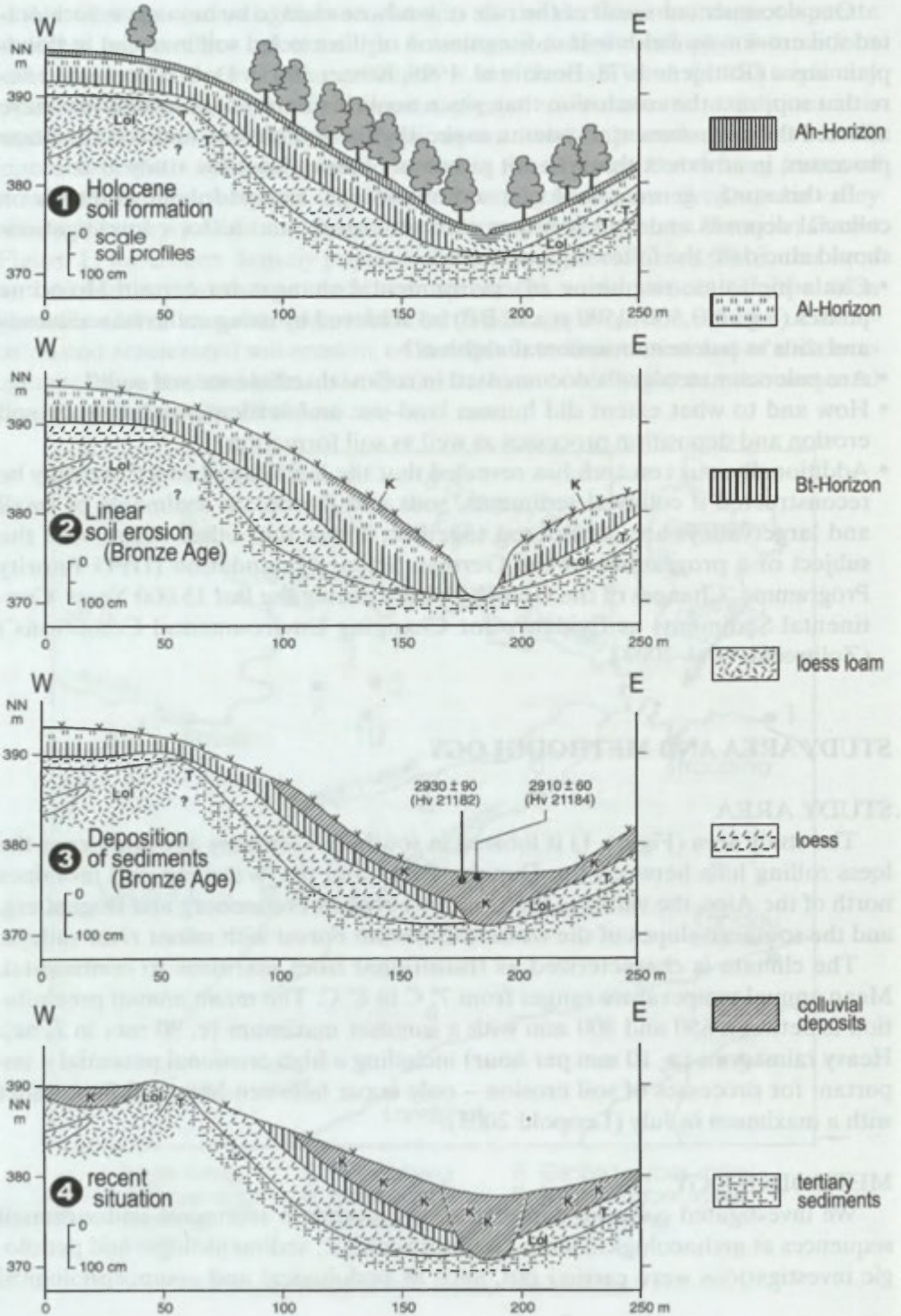
STUDY AREA

The study area (Figure 1) is located in southern Germany and comprises the loess rolling hills between the Danube River and the Würmian end moraines north of the Alps, the valley of the Danube between Frauenberg and Bogenberg, and the southern slopes of the western Bavarian Forest with minor river valleys.

The climate is characterized as transitional from maritime to continental. Mean annual temperature ranges from 7° C to 8° C. The mean annual precipitation is between 650 and 800 mm with a summer maximum (c. 90 mm in June). Heavy rainstorms (> 10 mm per hour) including a high erosional potential – important for processes of soil erosion – only occur between May and September with a maximum in July (Leopold 2003).

METHODOLOGY

We investigated colluvial deposits, soils, floodplain sediments and sediment sequences at archaeological sites. Geomorphologic, sedimentologic and pedologic investigations were carried out, such as pedological and geomorphological



mapping, descriptions and sampling of exposures (e.g., pipeline and road cuts) together with 1550 manual and automatic drillings. Soil physics and soil mineralogy (grain size, clay minerals, iron oxides, soil colour) and soil chemistry (organic matter, carbonates, soil acidity) were analysed in detail. Archaeological, ^{14}C age determinations of organic material (charcoal, wood, fossil organic soil horizons, bones) and infrared stimulated luminescence ages (IRSL) provided a chronological framework of Holocene landscape activity phases (soil erosion) and stability phases (soil formation). An overview of the methods and analysis used are presented by Niller (1998).

RESULTS

COLLUVIAL DEPOSITS AND SOILS AS ARCHIVES OF HIGH RESOLUTION CHRONOLOGIES

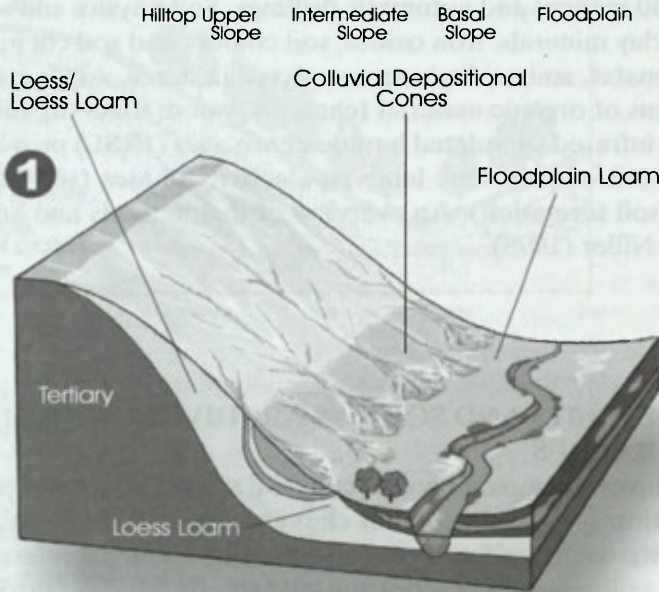
Our detailed investigations show that colluvial sediments cannot present high temporal resolution for environmental changes. Nevertheless, colluvial sediments and soils are useful geoarchives as combining elements intermediate between floodplain sediments and prehistoric land-use on adjacent slopes. The colluvial sequences of all investigated slopes reflect individual geomorphic processes in space and time. Prehistoric soil erosion occurred as rill wash and surface runoff. The periglacial relief, inherited from Würmian glacial times, was characterized by dells, small shallow valleys devoid of streams in the Holocene. These dells were traps for sediments transported downslope by sheet and rill erosion after anthropogenic clearance of the natural woodlands. Pronounced linear erosion processes, e.g., during the Urnfield Period at about 3,000 yr cal. BP, yielded short-term relief forms (steep rills and gullies) (Figure 2).

These rills and gullies document rainstorms of high erosivity within the prehistoric soil erosion system but any concentration in time, or a phase-like occurrence of these processes, cannot be proved. On slopes where a combination of sheet and rill erosion was active over longer periods, the accelerated soil erosion caused a remarkable reduction of the relief energy during the middle and late Holocene, because of colluvial sedimentation in Pleistocene dells. Furthermore, after the dells were filled with eroded material, colluvial sediment was deposited at the basal slopes and formed depositional cones (Figure 3).

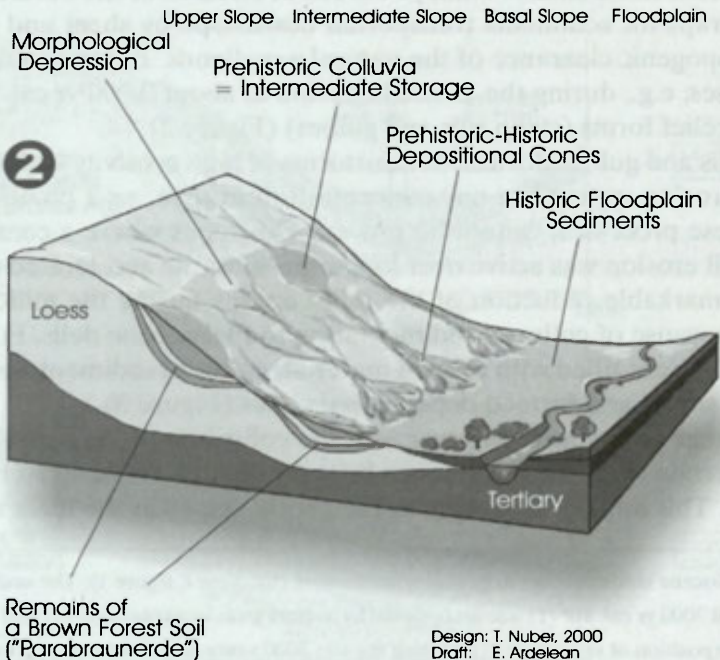
At this stage of soil erosion processes, the colluvium did not reach the floodplains. At certain sites, these processes brought about a decisive alteration of the slope relief. This applies especially to the gentle slopes in the loess rolling hills,

Figure 2. Holocene landscape development near Seedorf (location 7, figure 1): The undisturbed soil formation until 3000 yr cal. BP (1) was interrupted by a short term linear erosion (2) and an immediate subsequent deposition of sediments (3). During the last 3000 years only alternating processes of weak erosion and sedimentation occurred (4). No major linear erosion in the Middle Ages took place.

Step Slope in the Foothills of the Bavarian Forest



Gentle Slope in the Loess Rolling Hills



where many dells existed. In the foothills of the Bavarian Forest, the steeper slopes only show colluvial depositional cones at the lower parts of the slopes. Depending on variables such as vegetation, inclination, soil type, bedrock, geomorphology (dells, gullies), human impact (beginning, onset and end as well as intensity of agricultural activities, phases of settlements, land abandonment and population migration), the geomorphodynamic processes vary not only from slope to slope but also from time to time. Our investigations show that colluvial sediments have a limited value for paleoenvironmental reconstructions in space and time. We cannot transfer observations from one site to another site. This holds true for processes of soil erosion as well as for erosion phases in space and in time. Yet, when interpreted together with other geoarchives (lake sediments, pollen sequences and floodplain sediments), colluvial sediments and soils allow a detailed reconstruction of the paleoenvironments (Niller 1998, 2001). The advantage of our investigation is that colluvial sediment sequences, soils and floodplain sediments are to be found nearly everywhere; this enables us to reconstruct the landscape development not only of a certain site but of larger areas.

COLLUVIAL SEDIMENTS AS GEOARCHIVES TO RECORD CLIMATIC 'SIGNALS'

Climatic events in the form of single and/or repeatedly occurring heavy rainstorms are extremely important with respect to relief evolution by soil erosion processes. These events can only play a significant role, if anthropogenic land-use changes provide favourable conditions for sheet and rill erosion. Anthropogenic influences are climate-independent. Therefore, climatic events and/or climatic anomalies causing soil erosion may not be related to 'natural conditions'. A distinction between anthropogenic induced geomorphodynamic processes (e.g., accelerated soil erosion) and processes caused only by climate cannot be made. Colluvial sediments, standing by themselves, are not suitable to record climatic 'signals' (Leopold and Volkel 2002).

RELATIONS BETWEEN COLLUVIAL DEPOSITS AND SETTLEMENT HISTORY

During the middle and late Holocene, the most important type of relief development was the transformation of the natural landscape into a cultural landscape.

←

Figure 3. Landscape development in different parts of the investigation area according to a cascade model (Chorley et al. 1984). The Pleistocene relief controls the deposition of sediments (colluvial sediments, floodplain sediments): on the steep foothills of the Lower Bavarian Forest the transport to the basal slopes and into the floodplain dominates (1), whereas on the gentle slopes of the loess rolling hills with minor dells as sediment traps intermediate storage of colluvial deposits is more important (2). Hence, it is obvious that there are different time gaps between the soil erosion on the hills and the sedimentation of soil material in the floodplain areas. The duration of the time gap depends on the inclination and morphology of the Pleistocene and Holocene landscape.

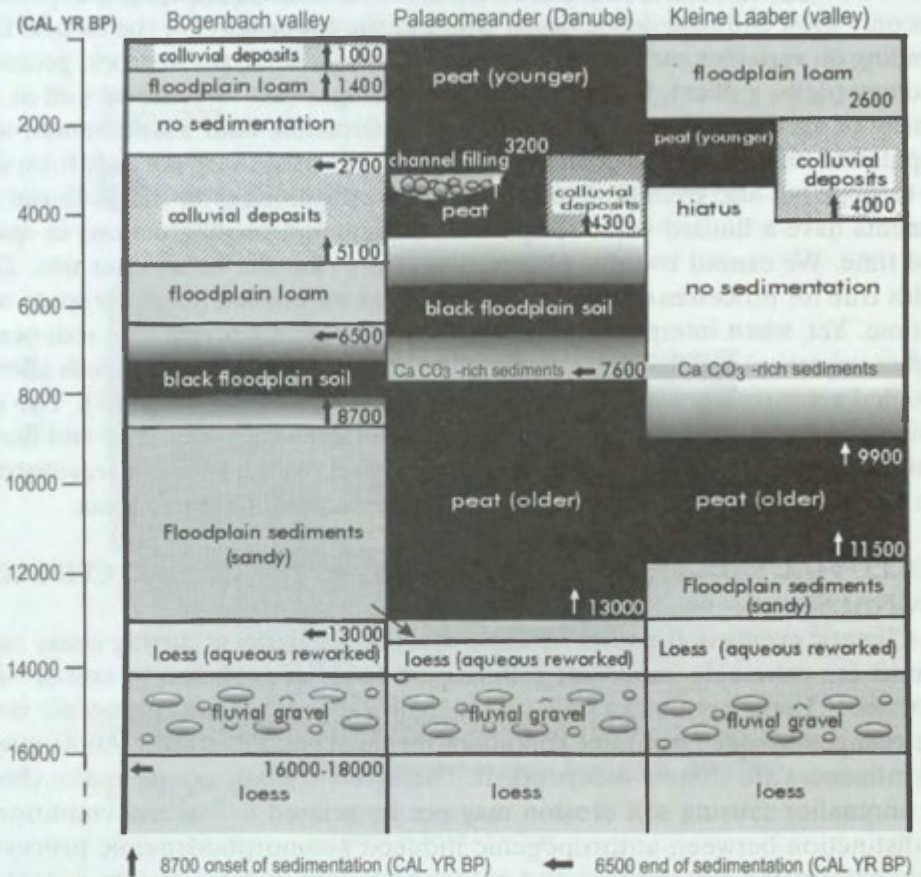


Figure 4. Chronostratigraphical framework of valley sediments in the loess rolling hills (Kleine Laaber, location 9, figure 1) and in the foothills of the Bavarian Forest (Bogenberg and Bogenbach, location 1, figure 1) compared with the stratigraphy of a paleomeander of the Danube (close to location 1).

pe (Heine 1994, 2000, Heine et al. 1997, Lang et al. 2003, Niller et al. 1997, Niller 1998, 2001). An important steering factor with respect to the result of the erosion and sedimentation processes is given by the late Pleistocene precolluvial relief with its periglacial forms. The morphodynamic 'cascade system' (Figure 3) describes the Holocene landscape changes. Colluvial sediments can be much older than floodplain sediments; Niller (1998, 2001) reports about a time gap between forest clearance by humans and the onset of accumulation of eroded soil material in the floodplain areas of the smaller valleys of up to several centuries and even of some millennia. The pre-existing Pleistocene relief controls the

deposition of colluvial and floodplain sediments. The time delay between the accumulation on the slopes and in the valleys depends on the characteristic relief configuration. Therefore, floodplain sediments alone cannot be used as geoarchives to explain the age, the nature and the extent of prehistoric soil erosion. By no means, the chronology of the settlement and land use history of the investigated areas is documented in the floodplain sediment sequences. Furthermore, recent studies show that the age of medieval colluvial sediments associated with anthropogenic land-use changes are decades or even centuries older than the data of written documents from archives referring to the settlement progress (Scheibe 2003).

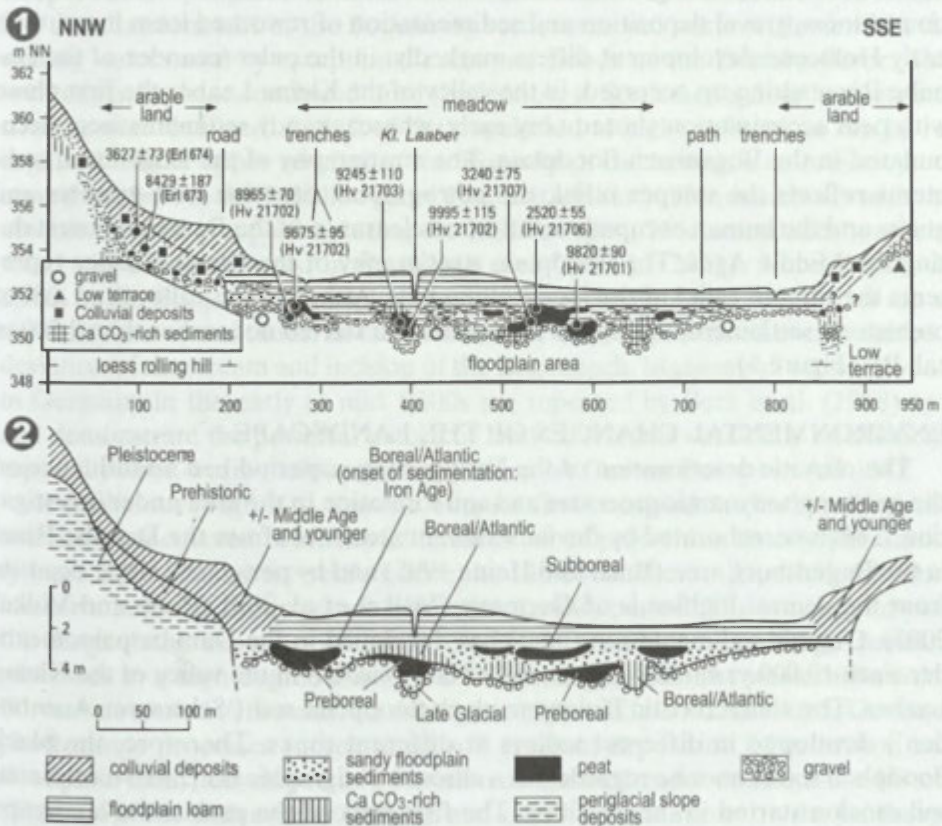


Figure 5. Schematic cross-section of the 'Kleine Laaber valley' (location 9, figure 1).

The upper cross-section (1) shows the detailed description of the sediments in the floodplain and on the hills. The lower cross-section (2) presents a synoptic view. At this site of the 'Kleine Laaber valley' the valley sediments are characterized by (base to top): fluvial gravels (Late Glacial) – aquatic reworked loess (Late Glacial) – sandy floodplain sediments (Late Glacial / Early Holocene) – older peat (Preboreal) – CaCO₃-rich sediments (Atlantic/Boreal) – younger peat (Subboreal) – floodplain loam (Subatlantic).

RELATIONS BETWEEN COLLUVIAL DEPOSITS, SOILS AND FLOODPLAIN SEDIMENTS

Only a synoptic analysis of colluvial deposits, soils and floodplain sediments provides a valuable reconstruction of the paleoenvironmental changes during the Holocene (Figure 3). Figure 4 shows the late Pleistocene and Holocene sedimentologic chronostratigraphies of three valley floodplains. The Bogenbach catchment lies in the relative steep relief of the Bavarian Forest; the catchment of the Kleine Laaber is situated in the loess rolling hills; the paleomeander of the Danube River lies near the boundary between the Holocene floodplain and the Rissian loess covered high terrace ('Hochterrasse'). Although the development of all three rivers during the late Pleistocene is much the same, with accumulation of loess, gravel deposition and sedimentation of reworked loess loam, their early Holocene development differs markedly: in the paleomeander of the Danube River silting-up occurred; in the valley of the Kleine Laaber the first phase with peat accumulation started very early, whereas sandy sediments were accumulated in the Bogenbach floodplain. The stratigraphy of the Bogenbach sediments reflects the steeper relief, the petrographic situation with granites and gneiss and the human occupation and wood clearance of the Bavarian Forest during the Middle Ages. The floodplain stratigraphy of the Kleine Laaber represents the smooth relief of the loess rolling hills. Although the valley is an axis of prehistoric settlement, floodplain sedimentation started not earlier than 2,500 yr cal. BP (Figure 5).

ENVIRONMENTAL CHANGES OF THE LANDSCAPE

The climatic deterioration of the Younger Dryas period had no influence on the geomorphodynamic processes and soil formation in the area under investigation. This is corroborated by fluvial sediment sequences from the Danube River in the Regensburg area (Buch and Heine 1995) and by periglacial slope deposits from the central highlands of Germany (Völkel et al. 2001, Raab and Völkel 2003). Organic sedimentation (peat) has developed in the Danube paleomeander since 13,000 yr cal. BP and since 11,500 yr cal. BP in the valley of the Kleine Laaber. The characteristic Holocene black floodplain soil ('Schwarzer Auenboden') developed in different valleys at different times. Therefore, the black floodplain soil cannot be regarded as a chronostratigraphic horizon. Accelerated soil erosion started in the neolithic. The first peak of the prehistoric landscape change occurred at the end of the Bronze Age (3,400–2,800 yr cal. BP). The first record of floodplain loam sedimentation (Figure 4) can be dated between 6,500 and 5,100 yr cal. BP in the Bogenbach valley and younger than 2,500 yr cal. BP in the valley of the Kleine Laaber. Exceptional rain events are recorded in sediment sequences and geomorphic features at the end of the Bronze Age and in Middle Ages (AD 1,100–1,400).

CONCLUDING DISCUSSION

There are many interactive relationships between human occupation and natural environment in the study area (Figure 6). The first settlements are located on flood-free, loess covered edges of fluvial terraces of the Danube at the boundary to the lower Würmian terraces that are covered by floodplain sediments. Until the Bronze Age these settlements were inhabited without major interruptions. During the Early and Middle Bronze Age, flood events diminished. Elevated sites of the lower terrace were settled. During the late Middle Bronze Age, the occupation of the Bogenberg and some other areas north of the Danube in steeper relief took place. Intensive soil erosion followed immediately. Presumably, the abandonment of the Bogenberg settlement and some settlements in the vicinity was caused by exceptional rains and floods of the Danube River. This 'wet' climatic phase continued until the Iron Age. Settlements near hamlets and streams were deserted. A shift of the population to higher elevations can be recognised. In the upper reaches of the Bogenbach catchment, a colonisation occurred not before the Middle Ages ('Hochmittelalter'). Simultaneously with the ubiquitous land occupation during the Middle Ages ('Hochmittelalter'), accelerated soil erosion started in the whole area (Scheibe 2003). Since about AD 1,100, in the Bogenbach floodplain area loamy sediments accumulated on top of the valley gravel and sands. In the 14th century catastrophic flood events caused a deviation of the stream and incision of the Bogenbach. Massive soil erosion events in Germany in the early to mid 1,300s are reported by Bork et al. (1998) too, and demonstrate the powerful nature of land-use/climate variability interactions, especially when a rare (1,000 yr) rainfall event hit Central Europe in AD 1342.

It is obvious (Figure 6) that the humans first occupied the loess rolling hills and the loess covered older fluvial terraces (since paleolithic times). The Tertiary loess rolling hills were settled since the Linienbandkeramik period, yet more intensive land-use occurred not before the Latène period. The area at and near the Bogenberg experienced anthropogenically induced landscape forming processes to a high degree during the Bronze Age, while the western Bavarian Forest was settled in the end, viz during the high and late Middle Ages. Geomorphodynamic processes (accelerated soil erosion, deposition of colluvium and floodplain loam) developed as a function of the settlement and land-use history (Figure 6). The influence of climate and climate fluctuations on the landscape changes was insignificant. If ever climate played a role, climatic consequences can only be explained in conjunction with the anthropogenic influences on the environment. The anthropogenic disturbances of the landscape (clearcuts, land-use) alone are responsible for accelerated soil erosion, accumulation of colluvial slope deposits, floodplain sedimentation, sheet erosion and gully cutting. Thus, humans are the driving force to shape their environment, since the neolithic until today, by purpose and/or unintentionally. Climate only modifies the processes caused by humans.

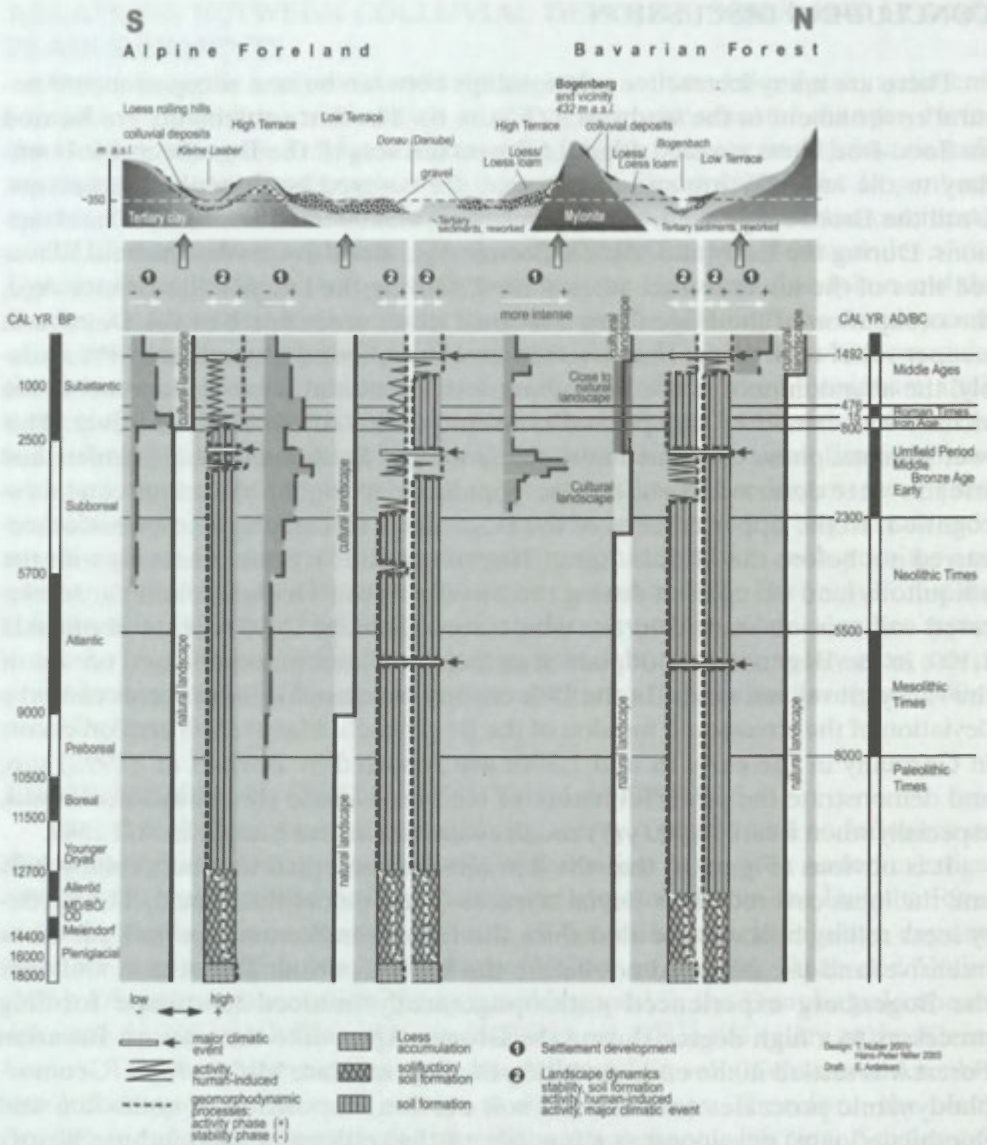


Figure 6. Landscape dynamics, settlement development, soil formation and climatic deterioration and their interactive relations in different parts of the study area.

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POSTGLACIAL EXTREME EVENTS AND HUMAN ACTION IN THE TRANSFORMATION OF ESTONIAN TOPOGRAPHY AND LANDSCAPES

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ABSTRACT: Knowledge of rapid changes taking place in natural environments is of fundamental importance for better understanding of man-induced processes, which should be recognized and managed. In Estonia, extreme events have been caused by meteorite explosions, earthquakes, and neotectonic processes, heavy storms, karst phenomena and human impact. Investigation of past processes provides a major key in predicting the changes that could be expected in the same or similar areas in the future. All the above-mentioned phenomena have been analysed on a local and regional scale, but they all are part of global-scale processes activated during the last decades.

KEY WORDS: extraterrestrial phenomena, earthquakes, land upheaval, floods, aeolian processes, karst, human impact.

INTRODUCTION

During recent decades the logical-scientific method of inquiry has advanced our knowledge of the natural world via independent observations, experimentation and precision measurement. One of the focal points of our studies is examination of the relationship between man and nature. In this paper different natural extraterrestrial, endo- and exogenic processes in the formation of environment will be discussed. But changes caused by human activity are often much stronger and, therefore, for predicting the changes in our surroundings we will briefly analyse also human impact, especially for the reason that human action can innovate, stop or change the direction of natural processes.



Figure 1. Location map.

The obvious rapid changes in Estonia have been rather well studied, particularly over recent decades (Figure 1). They affect the environment on daily, seasonal, decadal, and longer-time scales and are important from both the geological and social points of view, because local people have had to adapt themselves to the rapidly changing environment. Rapid changes in natural environments are also of fundamental importance for better understanding of human-induced processes, which need to be recognized and managed.

IMPACT EVENTS

The role of impacts of large asteroids and comets in the geological and biological evolution of Earth has been a subject of much recent debate. Several tens of impact craters, ranging from Neoproterozoic to Holocene in age, and from less than 100 m to more than 50 km in diameter, occur in both the crystalline basement and the sedimentary cover of the old East European Craton in the Fennoscandian—Baltic Region (Puura et al. 1994). The density of extraterrestrial phenomena is highest in Estonia where impact craters have attracted the attention of scientists since the beginning of the 19th century. In Estonia, 15 impact craters are known, and five meteorite falls have been registered during the course of the last two centuries. The size of Holocene craters ranges from 9 (Simuna) to 110 m (Kaali main crater). The meteorites that formed these craters must have been rather small.

The minimum size of a cosmic body that can cause a global catastrophe should be several kilometres in diameter. The diameters of the cosmic bodies

responsible for the formation of the Neugrund Crater east of the Island of Osmussaar in the Early Cambrian c. 535 million years ago, and Kärkla Crater on the Island of Hiiumaa in the Late Ordovician about 455 million years ago (Figure 1) are estimated as c. 400 and 200 m, respectively (K. Suuroja, personal communication 2001). Undoubtedly, the fall of those meteorites had a great regional influence.

Meteorites with a mass of several hundreds or thousands of tonnes can also cause severe damage, if they fall on densely populated areas. To investigate possible environmental consequences of such cosmic bodies, we recently examined the record of Holocene meteorite falls in Estonia and the likely imprint on the memories of local people (Raukas 2002).

In this paper, we will restrict analysis to the formative processes of the Kaali main crater on the Island of Saaremaa (Figure 1) where a lot of geological and geophysical investigations have been carried out for more than 70 years (Tiirmaa 1994). Veski et al. (2001) believe that the impact of meteorite explosion on the surroundings at Kaali must have been between that of the Hiroshima and Tunguska events. It must have induced wildfire which reached at least 6 km from the epicenter, i.e., to the Piila bog 6 km northwest of Kaali. The whole bog probably suffered from severe burn, judging from the extent of charcoal and wood layer. The burn brought about a change in peat accumulation and a shift from a *Phragmites–Carex–fern–wood* fen peat to a *Sphagnum–Eriophorum* ombrotrophic bog peat. According to Veski et al. (2001), the explosion also felled forests and the resulting cleared landscape is shown by the presence of wood stumps in the Piila bog, a decrease in tree pollen influx, and a greater influx of stomata in pollen samples. The impact explosion swept the surroundings clean of forest shown by the threefold decrease in the total pollen influx and the relative dominance of *Pinus* on the percentage diagram. The disappearance of cereals in the pollen suggests that farming, cultivation and possible human habitation in the region ceased for a long period.

Such conclusions may be based on incorrect estimation of the age of the impact and explosion energy of Kaali meteorite. Veski et al. (2001) compared the Kaali impact with the Hiroshima atomic bomb (15–20 kilotons of TNT) and Tunguska event (15 megatons of TNT). Probably, the Tunguska explosion took place at an altitude of 5–10 km above the Earth's surface. As a result, the Tunguska shock wave reached the ground producing a 4.5–5 magnitude earthquake, equivalent to 5–32 kilotons of TNT, therefore comparable to the Kaali impact. The Tunguska explosion devastated c. 2100 km² of forest and produced a radial burn of flora for more than 100 km. In Hiroshima, almost all life and structures up to 1.5 km from the epicenter were wiped out, collapse of buildings was observed up to 8 km away and flash ignition of dry combustible material was observed as far as 3 km from the epicenter.

Both in Tunguska and Hiroshima, the explosion was initiated in the atmosphere, but at Kaali most of energy was used for forming the crater and crater

mound (Figure 2) and only a small part of it was directed horizontally. Also the size of the meteorite (less than 3 m according to Reinvaldt (1933) and about 4.8 m according to Pokrovski (1963)) was too small to cause any ecological catastrophe in the environment. Therefore, we can conclude that in postglacial time meteorites produced only exotic hollows in the topography which should be protected as part of the natural heritage.



Figure 2. Dislocated and uplifted dolomite blocks in the wall of the Kaali main crater.

Photo by R. Tiirmaa.

EARTHQUAKES AND NEOTECTONIC MOVEMENTS

Earthquakes have also played a certain role in shaping the Estonian landscapes. According to Sildvee and Vaher (1995), 24 small (magnitude <math><3.5</math>) instrumental events with the intensity of III to VII, magnitude 1.5 to 4 and the depth of the focus 5 to 14 km, have occurred in Estonia between 1602 and 1992, causing slides and remarkable changes in the up-to-56-m-high coastal escarpments, cut in Ordovician (North Estonia) and Silurian (West Estonia) limestones.

The Osmussaar earthquake (Figure 1) on 25 October 1976, was the strongest one (magnitude 4.7, intensity VI to VII), affecting an area of 191,000 km², with an epicenter supposedly related to a fault zone running from the Central Baltic via the Island of Hiiumaa to the coast of Finland. Because of this earthquake, particularly large limestone blocks fell in front of the steep escarpments on the Island of Osmussaar.

The role of tectonics in the rapid change of topography has been insufficiently studied. Several, in principle contradictory, opinions have been expressed on

this matter. Some authors (Raukas et al. 1988) maintain that the late- and post-glacial uplift and the old Pleistocene movements of the Earth's crust, which occurred in Estonia under platform conditions, were for the most part isostatic and that tectonic movements were of little significance in morphogenesis. Others (Heinsalu and Sildvee 1971; Rahni 1973) state, in contrast, that many contemporary relief forms (e.g., eskers) are represented by morphostructures resulting from inherited differential movements of independent basement blocks.

Evidence obtained by analyzing the internal structure of the basement and bedrock, the thickness and facies of sedimentary cover complexes, and maps of the structure and contours of different key beds suggests (Raukas et al. 1988) that the tectonic framework has changed markedly and rapidly with time, and that the block structure of the basement is weakly reflected in the bedrock and contemporary topography. Linear elements, such as faults in the basement are acknowledged to be the most active and long-lasting forms, but even they are only modestly reflected in the bedrock and contemporary topography. The Haanja Heights (Figure 1), the highest region in Estonia, are situated on the Haanja-Lokno uplift block of the basement. Although tectonically active, the small amplitudes, low rate of neotectonic uplift and relatively thick Quaternary cover on the Haanja Heights show that tectonic movements have not been decisive in the formation of Quaternary deposits on these heights (Vaher et al. 1980).

During the late glacial and the Holocene, environmental evolution in Estonia was highly influenced by the glacioisostatic uplift of the Earth's crust, a process which is still in progress. Thus, in the 13th century, a big portion of the present Tallinn harbour was under the sea, with water reaching as far as the ancient town wall. At the site of the present Viru hotel in the central part of Tallinn, the water depth was at least 1.5 m. According to Kessel and Miidel (1973), in the course of 2450 years, since the existence of the Baltic glacial lake G_1 up to the formation of the Preboreal Yoldia Sea (Y_{II}), the territory of northwestern Estonia rose 65 metres. In the postglacial, which comprises the last 9,600 years, the same area rose about 50 metres. In the surroundings of Tallinn, the corresponding average rate of uplift during the above-mentioned period was 26.5 and 4.2 mm per year, respectively. At the present time this area is rising at a rate of 2 mm per year. As a result of the intensive uplift of the Earth's crust, the width of Estonia's coastal region exceeds 130 km and ancient coastal forms occur at various elevations (Kessel and Raukas 1979). Since the isostatic uplift in the northwestern part of the country was much more intensive than in the southwestern and northeastern areas, the synchronous coastal formations and the corresponding shorelines and old settlements there are situated at considerably higher altitudes than in the other parts of the country.

Land upheaval exerts a particularly great effect on the environment and on local people near the meridionally elongated lakes Peipsi and Võrtsjärv (Figure 1). For instance, at the end of the Late Pleistocene and the beginning of the Early Holocene the main outflow from Lake Võrtsjärv in Central Estonia was to the

west via the Tānassilma–Viljandi valley (Moora et al. 2002). As a result of the uneven isostatic uplift, which was more intensive in the north-west, this outflow was closed and the so-called Lake Big Võrtsjärv was formed. It covered large areas north of the current lake. At the beginning of the Middle Holocene an outflow was formed to the east via the Emajõgi valley into Lake Peipsi. As a result, the environment of the Mesolithic people around the lake changed rapidly and they had to adapt themselves to new living conditions.

At the present time, the northern part of Lake Peipsi is rising at a rate of 0.2–0.4 mm and the southern part is sinking 1.2 mm per year. However, in the past, the differences were greater. Those processes caused and are still causing the water to flow from north to south; as a result, large areas on the western bank lowlands and in the southernmost part of the lake (so-called Lake Pihkva), are severely flooded and/or clogged with muck. The most severe flooding takes place at the mouth of the Emajõgi River and the entire low shore of Lake Pihkva. In 1956, the flooded area reached 647 km² (Sokolov 1983). During the last 1,600 years, the water level rose 0.6–0.8 mm per year in the area of the estuary of the Emajõgi (Müidel 1981). According to R. Pirrus (Pirrus and Tassa 1981), the water level in Vāraska Bay (Figure 1) rose 10 m over the last 8,000 years. According to Mieler (1926), the area of Piirisaar Island in the southern part of Lake Peipsi proper was 20.08 km² in 1796 and 10.64 km² in 1834; at the present time it is only 7.39 km².

The gradual southward movement of the lake waters has led to the submergence of several Neolithic settlements in the vicinity of the lake (Jaanits et al. 1999). The chronicle of Pskov records that in 1458 a church was built on the Island of Ozolitsa. The remains of this church are now submerged beneath 0.4–2.5 m of water (Tyumina 1966), and two small islets (Lezhnitsa and Stanok) are all that remains of the former island. Tyumina (1966) points out that the inhabitants of many villages had to leave their homes and to retreat farther inland in front of the advancing lake waters. The remains of the old Chudskaya Rudnitsa Village now rest at a depth of 0.5–1 m in the lake, some 100–200 m offshore.

AEOLIAN PROCESSES AND COASTAL EROSION

In Estonia wind velocity may reach 40–45 m/s. Strong winds can cause severe damage to the environment. On several occasions in the past, the aeolian sand posed a great threat to inhabitants as well. At the end of the 19th century, mobile dunes on the Island of Saaremaa endangered Kärļa church, the pastor's mansion and farmsteads in the vicinity. The advance of the dunes was stopped by a pine stand (Tiismann 1924). The same author describes the movement of sand in the surroundings of the Ristna lighthouse where the strong southwesterly winds picked up a mass of sand and deposited it behind the doors. Every time the islanders had to work for days to remove the sand and clear the access to the houses.



Figure 3. Moving sands on the northern coast of Lake Peipsi on June 18, 1998.
Photo by A. Raukas.



Figure 4. The coast of Valgerand north of Parnu after a heavy storm on August 13, 1997.
Photo by A. Raukas.

At present, the movement of aeolian sands is limited (Figure 3) and has more or less been stopped by the planting of vegetation. At the same time, the intensification of agriculture and large-scale land improvement in the 1950s–1960s brought about huge fields with a steppe-like appearance and caused deflation on sandy and peat soils. There are some 200,000 ha of land endangered by deflation in Estonia. On Hiiumaa Island (Figure 1), such fields make up two-thirds of the arable lands. When the danger was understood, fields larger than 50–60 ha on the lowlands and 20–30 ha on more elevated areas were prohibited. Belts of trees were planted to shelter fields, while gentle peat soils were planned for a longer-term use as grasslands. As a result of the liquidation of state large-scale agriculture in Estonia, the area of fields endangered by deflation has essentially decreased during the last decade.

Heavy storms also cause severe damage to the coasts (Figure 4), especially in the autumn-winter period, when the water level is relatively high and persistent unidirectional winds pile up great water masses and the surge can reach several metres.



Figure 5. Ridges of pressure ice near Nina Village on the western shore of Lake Peipsi on March 20, 2002.

Photo by A. Miidel.

Rapid changes on the beaches are also related to ice-push action (Figure 5). Ridges of pressure ice, up to 10–15 m high pushed forward against the shore with enormous force, play a significant role in shaping the shore and transporting huge erratic boulders in the coastal area. Ice-push is at its greatest on sandy beaches at Pärnu Town (SW Estonia) where under the compressing and ploughing influence of the ice deep furrows are often formed on the beach and coastal slope. On till shores, furrows frequently occur behind the blocks moved by the thrust

of expanding ice. Occasionally, tens-of-metre-long stone walls are encountered in front of scarps which they defend from further erosion. During a milder and ice-free winter, the erosion of beaches is much more intensive. At the same time, the up-pressured sea and lake ice can cause extremely rapid changes in the coastal area, crushing the scarps and bluffs and piling on the beach up-to-two-metre-high and tens-of-metre-long accumulations of loose sediments.

KARST PHENOMENA

Over a half of Estonia's territory has a calcareous bedrock overlain by a thin Quaternary cover. In vast areas, so-called alvars, the Quaternary cover is missing. Therefore, karst topography and underground karst features resulting from the solution of rocks and leaching processes are common. Karren, karst holes, open jointings, karst relics, funnel-sinks of absorption, angular subsidence sinkholes, underground and disappearing rivers, caves and other karst phenomena have been identified in Estonia (Heinsalu 1977). In some places, peculiar temporary karst lakes occur. In the central part of Pandivere Upland (Figure 1), in an area of 1050 km², surface rivers are completely absent and practically all rain and snowfall goes via cracks and more than 200 karst holes below the earth surface. The water flows underground to the fringes of the upland, where it appears at the surface in the form of copious springs, which turn into rivers running in all directions. Karst landscapes with active sinkholes are highly vulnerable. The most serious hazards occur where ground collapse damages communication systems, agricultural fields, roads and buildings. Another serious problem is deterioration of groundwater quality and underground oil-shale mining. Every year in Estonia some new sinkholes appear which can cause not only obvious damage to different structures but open new pathways for pollution of groundwater. This is especially dangerous in urban areas, near sewer systems and landfills, contributing to the penetration of contaminated and chemically active water into the bedrock and groundwater aquifers.

RAPID CHANGES CAUSED BY HUMAN IMPACT

The territory of Estonia has been inhabited throughout the Holocene. Land cultivation did not begin to dominate the life of ancient people until the beginning of the Early Iron Age, about 600 years B.C. Since that time, man has inflicted incurable wounds on nature. The vigorous intensification of agriculture and industry during the Soviet occupation was accompanied by a sharp increase in the exploitation of mineral resources and by an ever worsening impact on the environment. During 50 years of occupation, the population of Estonia increased some 1.4 times, the number of workers and employers 3.8 times, industrial output 4.2 times, the production of mineral resources 15 times, and the generation of electric power 100 times.



Figure 6. Mining of oil shale in Narva open pit on June 11, 2002.

Photo by A. Raukas.

Human activities severely damaged the natural landscape in about 8 percent of Estonia's territory, and drawing up recommendations for land improvement is one of the most important tasks facing Estonian scientists today. This is not easy because annual exploitation in the mining industry alone amounts to tens of million tonnes of solid mineral resources, not including the sand used as a ballast material and the overburden removed from open pits (Figure 6). In 1991, the territory of the mined area increased by 344 ha in mines and by 269 ha in open pits (Paalme 1993). Every year about 600 ha of land were lost to oil-shale pits. In northeastern Estonia about 250 ha are covered by ash-fields and waste hills.

As a result of the large-scale land improvement carried out in the Republic during Soviet times the water table dropped in lakes, river runoff decreased, and several rivulets and springs dried up in the low-water period. In some areas natural river valleys are practically absent. For example, artificial canals and dredged river channels account for 77 percent of the total length of flowing waterways in the Kasari River drainage system (Figure 7).

About 30,000–40,000 ha of agricultural land and 15,000–20,000 ha of forest land were improved annually. Already by 1983, about one million ha of land had been drained (Aruja 1983). In Estonia the water table in 300 lakes (out of 1500) is regulated. About 150 water reservoirs, making up 1.5 percent of Estonia's ter-

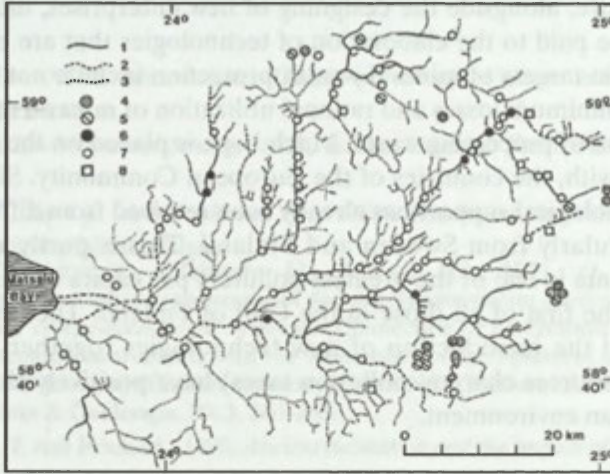


Figure 7. Human-induced changes in the catchment area of the Kasari River (after L. Veering, 1983): 1 – natural stream bed; 2 – regulated stream bed; 3 – artificial stream bed; 4 – lake; 5 – overgrown lake; 6 – artificial lake; 7 – former artificial lake; 8 – water reservoir.



Figure 8. Ruined Soviet missile base at Keila-Joa west of Tallinn on June 14, 1999.

Photo by A. Raukas.

ritory, have been constructed. Extremely great damage to the Estonian environment, estimated at about 4 billion US dollars (Raukas 1999), was caused by the Soviet Army (Figure 8). There were 1565 military objects of the former Soviet Union in Estonia, which occupied some 87,000 ha of land, i.e., 1.9 percent of Estonia's territory.

In the future, alongside the designing of new enterprises, much more attention should be paid to the elaboration of technologies that are environmentally safe. The main targets of mineral wealth protection include not only exhaustive mining with minimum losses and rational utilization of mineral resources, but also the disposal of processing waste. Much hope is placed on the support of, and cooperation with, the countries of the European Community. Significant financial and technological support has already been received from different European states, particularly from Sweden and Finland. This is partly understandable because Estonia is one of the greatest polluters per capita of European air and waters, and the first of all those in the Gulf of Finland. The decline in mining activities and the introduction of new technologies together with economic measures (resources charges, pollution taxes) have positively changed the state of the Estonian environment.

CONCLUSIONS

Investigations of rapid natural processes of the past provide a major key to understanding of on-going processes and assist in predicting the changes to be expected in the same or similar areas in the future. Technological developments were accompanied by a dramatic increase in human impact. Some changing environments in Estonia reflect global trends (warming of the climate, depletion of the ozone layer), whereas others are local or regional trends, which, nevertheless, are important to compare with global scale processes. It seems that the recent decades have witnessed a remarkable activation of erosion processes and the frequency of extremely heavy storms has increased. In 1996 the Estonian environmental strategy was compiled (Estonian... 1997). It determines the development of Estonian nature use and environmental protection in the new political and economical situation and specifies the priority goals in the use and protection of natural resources. In the strategy land use and landscape protection problems are those having prior claim for consideration. Prediction and management of rapid changes can help to guarantee maintaining a healthy and satisfactory living environment for humans.

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EXTREME EVENTS IN THE CONTEXT OF LATE QUATERNARY ENVIRONMENTAL CHANGE

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PART II PERSPECTIVES AND APPLICATIONS

ABSTRACT: The significance of extreme events in relation to change in landscape is the subject of the late Quaternary Millennial-scale cycles of climate change, the most recent of which have been the most dramatic. However, the late Quaternary also witnessed the development of a number of landscape features and forms which were parallel to, or subsequent to, these cycles. This paper reviews the evidence for extreme events in the late Quaternary and discusses the implications of these events for the development of landscape features and forms. The implications of extreme events for the development of landscape features and forms are discussed in the context of the late Quaternary environmental changes. The implications of extreme events for the development of landscape features and forms are discussed in the context of the late Quaternary environmental changes.

KEY WORDS: climate change, extreme events, landscape evolution, late Quaternary

INTRODUCTION

The impacts of extreme events on geomorphic systems are determined by the course of change over long periods during which severe events occur. But the relationship is far from straightforward. Starkel (1976) has pointed out, for example, that gullies can form rapidly during extreme rainfall events, and they alter the course of slope evolution. He also noted that progressive vertical aggradation was a long-term result of catastrophic surface erosion and led to a similar change in the slope profile. On the other hand, he recognised that newly-created features may be short-lived, if the initial energy impulse is not repeated. Starkel has also warned (1976, pp. 240-241) that "extreme processes of the present may be generally discordant with the trend of slope evolution inherited from past periods." This means we should be cautious not to refer them to long-term

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PART II
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EXTREME EVENTS IN THE CONTEXT OF LATE QUATERNARY ENVIRONMENTAL CHANGE

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ABSTRACT: The significance of extreme events in landform change is discussed in the context of the late Quaternary. Millennia-scale cycles of climate change are less than the relaxation times of most landform systems and have led to widespread disequilibrium in natural landscapes. Slope failures and fluvial systems show parallel evolutionary trends in high and low latitudes, resulting from late Quaternary environmental changes. The transformation of geomorphic systems appears to require millennia of preparation time and research needs to establish the role of individual events within this timescale of enquiry.

KEY WORDS: climate change, extreme events, fluvial systems, slope failures.

INTRODUCTION

The impacts of extreme events on geomorphic systems can determine the course of change over long periods during which lesser events occur. But this relationship is far from straightforward. Starkel (1976) has pointed out, for example, that gullies can form rapidly during extreme rainfall events, and may alter the course of slope evolution. He also noted that progressive upslope aggradation was a long-term result of catastrophic surface runoff and led to a secular change in the slope profile. On the other hand, he recognised that newly-created features may be short-lived, if the initial energy impulse is not repeated. Starkel has also warned (1976, pp. 240-241) that 'extreme processes of the present day are generally discordant with the trend of slope evolution inherited from past periods.' This means we should be 'cautious neither to refer them to long time-

spells nor to connect them with other climatic conditions' without supporting evidence. This essay seeks to examine the role of extreme events within a Quaternary timescale.

One way in which we have sought to understand the impact of system disturbance is to refer to the 'relaxation time' or period required to re-establish landscape equilibrium. However, several problems surround this concept, because geomorphic systems:

- may not have been in equilibrium prior to the timing of extreme events,
- may require periods of 10^3 - 10^5 , perhaps even 10^6 years, to re-establish equilibrium,
- will, therefore, experience '100 year' extreme events when equilibria are already disturbed,
- will undergo internal readjustments or 'complex response' to disturbance,
- are unlikely to return to their previous states because of irreversible changes,
- will adjust gradually to long-term climate changes over 10^3 - 10^5 year periods, which alter the controlling parameters, including the magnitude and frequency of 'extreme events',
- may be subject to differential rates of change in extrinsic factors, including rainfall, vegetation cover and soil properties,
- will evolve over long time periods as a consequence, inter alia, of base level control, continued rock decay, and downwearing of headwaters.

Exemplification of these problems can demonstrate some important realities concerning stability and sensitivity to change in natural landscapes.

The Quaternary Era provides a useful time frame, because its 10^6 yr duration encompasses major landscape changes, while the operation of Milankovitch and sub-Milankovich cycles of change provide time spectra (10^3 - 10^5 yr) against which the occurrence of landforming events can be measured. In practice, we are largely confined to the last glacial cycle (LGC) of 125 thousand years for our analyses, and much published research has been limited by the range of radiocarbon dating. But increasing use of luminescence techniques for dating quartz sand grains, and the dating of tephra, speleothems and ocean sediments using other radiometric techniques, is extending the range of our understanding rapidly. As a consequence, observations of geomorphic systems and their responses to extreme events within the period of record (10^2 yr) can now be placed, albeit imperfectly, within longer time frames.

These longer time periods also raise important conceptual issues, including the use and meaning of terms such as 'abrupt' or 'rapid' change. When employed in relation to short-term records (10^1 - 10^2 yr) these terms denote changes taking place in response to specific events, and attempts may be made to construct an 'event stratigraphy'. But, when viewed across the 10^4 - 10^5 yr time frame of the Late Quaternary, rapid change will usually refer to transformations taking place over 10^2 - 10^3 yr (Thomas 2004). This millennial scale of interpretation has been specifically recognised in some recent studies of European rivers (Veldkamp and

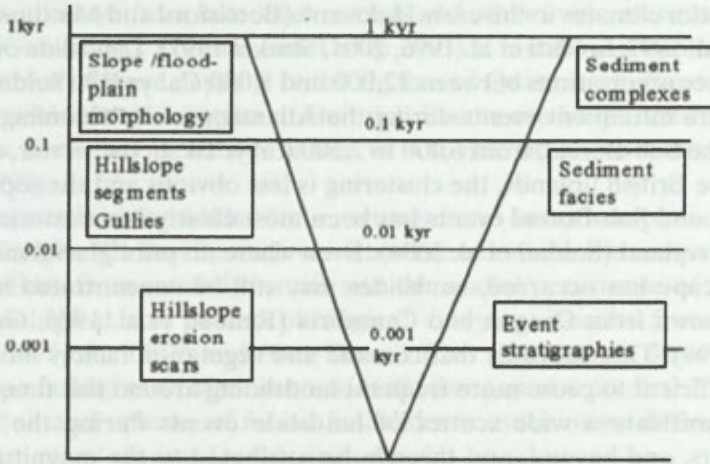


Figure 1. Perception of 'rapid' change in relation to the timescale of enquiry.

Over periods of 0.001 k yr (1 yr) focus is on specific erosion and deposition events. When the time scale is increased to 1 k yr, the reorganization of hillslope, piedmont and alluvial systems can be recognized.

Tebbens 2001, Lewis et al. 2001), and it also emerges from detailed studies of the R. Vistula (Starkel 1995b, Kalicki 1996). Although the sampling and dating may attain a century-scale resolution, there are good reasons to consider the millennial scale of enquiry important. This is, first, because there is evidence for millennial-scale sub-Milankovich cycles of climatic fluctuation in the Quaternary record (Bond et al. 1997) and, second, because fluvial transformation in response to environmental changes may operate on a similar time scale (Vandenbergh and Maddy 2001). In the case of the Vistula, Kalicki (1996) is clear that the evidence shows that the Holocene evolution is connected to climatic rather than anthropogenic changes. In many areas this question remains a matter of debate, and some forms of direct human intervention cause system changes on much shorter timescales (bedload traps, channel diversion, rapid deforestation). This question will not be pursued here. But it can be seen that our understanding of landscape response to system changes is relative to the scale of enquiry (Figure 1).

ILLUSTRATIONS AND PROBLEMS

SLOPE FAILURES

Slope failures (landslides) provide valuable insights into landscape evolution, because they are discrete events that create well-defined morphologies. In high latitudes there is increasing recognition that landslide events tend to be clustered in time and were particularly frequent following the final withdrawal of glaciers after the Last Glacial Maximum (LGM). In some cases this is a paraglacial effect, due to glacial over-steepening, unloading, melting of ground ice, and the

impacts of wetter climates in the early Holocene (Berrisford and Matthews 1997, Matthews et al. 1997, Soldati et al. 1996, 2004, Starkel 1997). Landslide events in Europe show concentrations between 12,000 and 8,000 Cal yr BP (Soldati et al. 2004), but were infrequent events during the Atlantic period, becoming clustered again in the Sub-Boreal, from 6,000 to 2,500 Cal yr BP. In the wetter, oceanic climates of the British uplands, the clustering is less obvious and the separation of the Boreal and Sub-Boreal events has been most clearly demonstrated from Italy and Switzerland (Soldati et al. 2004). Even where no prior glacial modification of landscape has occurred, landslides may still be concentrated in these periods, as shown from Oregon and Cantabria (Reneau et al. 1986, Gonzalez Dias et al. 1996). This indicates that climatic and vegetation factors alone may have been sufficient to cause more frequent landsliding around this time. Other records demonstrate a wide scatter of landslide events during the last 20 thousand years, and beyond, and this can be attributed to the magnitude and frequency of triggering events.

Dating of landslides in tropical environments has not so far provided unambiguous evidence for similar clustering. One reason for this is the paucity of data; another is the important fact that there was no 'glacial interruption' to the evolution of slopes during the Quaternary. Hillslope deposits from 'pre-glacial' times are, therefore, likely to persist in the landscape. This latter point is of great significance and not simply because a wider scatter of potential landslide ages is theoretically possible. The existence in the landscape of landslides of widely varying ages creates a complex mosaic of palaeoforms and deposits, some of which become weathered and indurated, limiting the potential for further large-scale movement. The importance of late Quaternary landslide events has, however, been demonstrated from Eastern Brazil (Modenesi 1988; Modenesi-Gauttieri 2000), where three major phases of deposition on the slopes of the Campos do Jordão Plateau were reported: 37-31 thousand years BP, 22-14 thousand years BP, 9-8.5 thousand years BP. The last two of these periods bridge the LGM and the early Holocene 'pluvial' respectively. Shroder (1976) mapped more than 200 landslides in southern Malawi but was able to obtain only one date of around 12 thousand years.

In Eastern Zambia, Central Africa, in a savanna climate with c. 1100 mm per year of precipitation, a series of large landslides was mapped around a series of deeply weathered residual hills (Thomas 1999; Thomas and Murray 2001a). Initially, it was expected that these very fresh forms might date to the period of climate warming after the LGM. But OSL dates taken from intercalated sediments show that some of the major slides are >200,000 OSL yr in age, and, because all the slides have similar morphologies, it is possible that the entire group of slides dates to the middle Quaternary. A later generation of smaller slides remains undated. Other colluvial and alluvial sediments in the area indicate a continuous process of accumulation throughout the Last Glacial Cycle (LGC). But because Quaternary climatic fluctuations in Eastern and Central

Africa are known from lake level records and other evidence to have been profound (Gasse 2000, Thomas and Thorp 2003) it is still reasonable to search for explanations of these deposits in the rhythm of Quaternary climate change. It is even possible that the impact of Heinrich Events can be detected in tropical sequences, since they have been shown to influence off-shore sedimentation (Arz et al. 1999).

In areas where landslides are a recurrent and hazardous phenomenon, as in Hong Kong or SE Brazil, much work has been done on the temporal prediction of landslides from rainfall records and the spatial occurrence of slope failures from geological and geomorphological analyses. Results from Hong Kong (Lumb 1975, Au 1993) and from Puerto Rico (Larsen and Simon 1993) show that slope failure as a response to rainfall events can be predicted. But the actual location and volume of future landslides is much more difficult to predict. There are many reasons for this, including:

- effectively stochastic variations in storm intensity across complex terrain,
- unequal inherent liability for sensitive slopes to fail, which may subsume:
- variation in time elapsed since last failure at different locations,
- hidden structures and fracture patterns,
- existence of unmapped older landslides.

With regard to this last factor, in many areas recent landslides have often formed within the amphitheatres and on the residual slopes created by much older slope failures. In this situation prior events have left scars and deposits with very long relaxation times and re-adjustment of the slopes is an ongoing process after 10^5 yr. From personal observation this factor operates in Hong Kong, the Freetown Peninsula and many other mountain areas in Sierra Leone (Thomas 1998) and the question was raised by Nilsen and Turner (1975) with respect to California. In central Sierra Leone (rainfall c. 2500 mm per year), Gaskin (1975) recognised two major phases of landsliding in the Sula Mountains (Thomas 1994). These remain undated, but the widespread formation of ferricrete, in the deposits, suggests that they may be very old. Ferricreted hillslope breccia around another hill group, the Kasewe Hills, produced evidence of reversed polarity in the iron compounds, and this implies a period of chemical precipitation >700 thousand years (Thompson 1979).

In NE Queensland, ongoing work is revealing that older landslide lobes along the E-facing escarpment N of Cairns (17° S Lat.) have varying ages within the LGC, only one of three currently dated by OSL methods, having occurred shortly after the LGM. In each case there is evidence for renewed Holocene activity post 6 thousand years. This Holocene activity has generally led to debris flows following current stream beds, spewing coarse sediment on to lower slopes as a fan, or has taken the form of diffuse colluviation of fine sediment from the weathered metamorphic rocks (Nott et al. 2001, Thomas et al. 2001b).

One possibility that must be considered is that areas with large palaeolandslides have responded to neotectonics. In coastal areas, seismic shocks might have

occurred as a consequence of eustatic and isostatic changes following the LGM (Thomas, 1998). In the Sierra de Mantiquera of Eastern Brazil, Modenesi-Gautieri (2000) also documents evidence for Quaternary neotectonics. In Eastern Zambia, the landslides are found in an apparently stable cratonic landscape remote from the sea (Thomas 1999), but their probable age of $>200,000$ would date them to a period of possible movements on the nearby Malawi Rift (Flannery and Rosendahl 1990). Any such association, however, would be highly speculative, and the slides give evidence for some fluidisation of the debris, which has produced lobes of debris extending >0.5 km from the slope base. However, such observations can also be applied to areas of known tectonism such as the Japanese Alps and Taiwan.

Fundamental to all landslide analysis is the relationship (ratio) of shear strength to shear stress, which can be defined as the 'factor of safety'. Shear strength is dominantly influenced by the properties of materials (such as clays formed by rock weathering) and the discontinuities within them. Many weathered slopes in the tropics approach a factor safety of 1, which forms a threshold below which slopes are liable to fail (Thomas 1994). Simultaneous increase in shear stress and reduction of shear strength tends to result from prolonged intense rainfall, and mainly shallow slope failures can be predicted on this basis. But these concepts are less easily applied to deep-seated failures, where repeated movement along the shear plane may be more influenced by prolonged periods of increased wet-ness such as would result from decadal to millennial shifts in climate parameters.

A number of conclusions might be drawn from these limited observations:

- in the absence of specific structural or tectonic controls, the occurrence of large, deep seated landslides may be related to major transitions or episodes of landscape change,
- the existence of older, deep seated slides in areas of recent shallow failures could indicate the past effects of neotectonics (earthquakes),
- in glaciated areas the withdrawal of glaciers at the last Termination marks the onset of a major landsliding event,
- in non-glaciated areas, although major hydrological changes took place at this time, older slides testify to previous extreme events or periods of climate transition favourable to slope failure. In the tropics it is tempting to speculate that these will have coincided with episodes of rapid climate change, perhaps of warming and increased rainfall, but evidence remains sparse,
- in landscapes that have experienced such episodes of Quaternary slope failure, subsequent landslides are frequently shallow and of smaller lateral dimensions,
- such large slides are unlikely to recur at the same site for a very long period of time because they have reset the parameters governing subsequent slope failures,
- recent slides in most areas tend either to continue the slow re-adjustment of postglacial slopes (paraglacial effects) or they respond to individual extreme events or short-term periods of wetness, in some instances the response to

extreme events of very high magnitude has been the occurrence of very large numbers of relatively small failures, not the reactivation of large slides.

This last observation is illustrated from Eastern Brazil, where an estimated 10,000 debris flows occurred in January 1968, in a small area of the Serra das Araras, in response to a single storm producing rainfall intensities of 114 mm per hour and a total 24 rainfall exceeding 300 mm (Jones 1973).

Many engineering studies refer to the need for 'ripening of slopes' following landslide events, and such a concept is built in to many studies of landslide recurrence intervals. Benda and Dunne (1978) indicated that the age of basal colluvium in bedrock hollows in Northwestern USA implied a recurrence interval of around 6000 yr. But when taking the whole landscape with its steeper slopes in 1st and 2nd order basins into account this interval fell to 750 yr and 1500 yr respectively. While details are scarce, these records and observations set both a context and limits to more conventional magnitude and frequency studies.

FLUVIAL SYSTEM CHANGES

Although the parameters are different, similar principles apply to fluvial systems and the ways in which they respond to extreme events and climate changes on different timescales. Fluvial systems are often portrayed as responsive to individual flood events and much research time has been spent on the documentation of fluvial changes in response to floods. Changes in bed morphology, channel avulsion, boulder deposition from debris flows all feature in these analyses, illustrated by detailed work from the Upper Vistula by Starkel (1998). However, in most cases, stream regimes revert to their previous habit: meandering, braided or other morphologies such as fan deposition being maintained. On the other hand it is equally evident from the Quaternary record that rivers have switched from braiding to meandering on quite different timescales, as reviewed for the temperate zone by Starkel (1987, 1995a,b). Furthermore, alluvial plains have complex depositional histories and, as with hillslopes, inherited morphology influences and limits continuing development.

The role of extreme events in the development of alluvial plains is, therefore, complex, and references to 'rapid' or 'abrupt' changes to fluvial systems require careful definition (Thomas 2003). Such terms only acquire meaning when the timescale of the enquiry is defined. In the context of Quaternary studies of river systems, 'rapid' change is usually defined at the millennial scale. The reasons for this are interesting, because although it is easy to dismiss this timescale as a reflection of data resolution from alluvial sections and, therefore, as an artifact of sampling and sample analysis, the situation is more complex. River response to decadal-scale changes in discharge regime were illustrated, for example, from Arizona by Hooke (1996) in respect of channel-pattern changes, but even here, major change in fluvial style within this timescale is not argued. The critical threshold conditions necessary for channel 'metamorphosis' are recognised as complex, and Schumm (1977) pointed out that channel characteristics (especially bar

patterns) vary greatly over time within a single channel. Changes to river discharge and bed-load can transform channel morphology on annual to decadal time-scales, depending on seasonal and inter-annual shifts in the controlling parameters. But the sensitivity of entire river systems to widespread transformation appears to depend on significant environmental change and the crossing of external thresholds (Werritty and Leys 2001).

Studies of the Upper Thames (Lewis et al., 2001) and rivers such as the Meuse and Vistula (Starkel 1995b, Kalicki 1996), together with modelling experiments (Veldkamp and Tebbins 2001) have all confirmed the importance of the millennial scale in revealing the most important transitions experienced by major river systems. In the tropics, few detailed studies have the resolution to prove or disprove this view, but the recorded changes all fit well into the millennial time frame. By this it is implied that important fluvial system changes, for example from braided to meandering channel patterns or from fan deposition to incision, appear to take place over millennia rather than shorter time periods. Vandenberghe and Maddy (2001), for example, asked whether the duration of the Younger Dryas (YD) interval in Europe (c.1100 yr) was long enough to evoke a fluvial response. The same might be asked of the records from West African rivers (Thomas and Thorp, 1995), where the cool dry conditions of the YD may have led to reduced flows and fewer sedimentary units, but induced no widespread system changes. During the pluvial conditions that followed the YD in Africa, silts from large floods were often deposited on older terraces of earlier sand- or gravel-bed rivers. Floods from the Nile both before and after the YD interval are recorded by sapropels in the Eastern Mediterranean (Rossignol-Strick et al. 1982). The global picture shows that ice cores and N Atlantic ocean sediments both record sub-Milankovitch cycles of around 1500 yr duration, otherwise known as Bond cycles (Bond et al. 1997). The timing of Heinrich Events and Dansgaard-Oeschger interstadial events confirms this analysis, and ocean records off NE Brazil demonstrate the relevance of these intervals to the understanding of terrigenous sediment fluxes to ocean sinks in tropical latitudes (Arz et al. 1998).

However, all events occupy a time continuum, and if that continuum involves a secular change over 10^3 yr, then extreme events with decadal frequencies will occur at different points along the curve of change. Two storm events of similar magnitude may, therefore, have different impacts according to their position along this curve (Figure 2). But, if this is the case, then it is necessary to offer reasons for such behaviour. The factors involved will either be intrinsic or extrinsic to the system. With fan deposition, for example, progressive steepening of gradient might tip the system towards incision. But this is unlikely if water flows remain episodic and channel load stays high. On the other hand, if the climate becomes wetter a series of changes may affect the system, including the growth of more dense vegetation, which enhances slope stability and promotes weathering and soil development. These changes will affect the shape of the hydrograph and pattern of sediment delivery, leading to more stable channel patterns and

possibly to incision. The question for this discussion is the time necessary for such extrinsic factors to become effective.

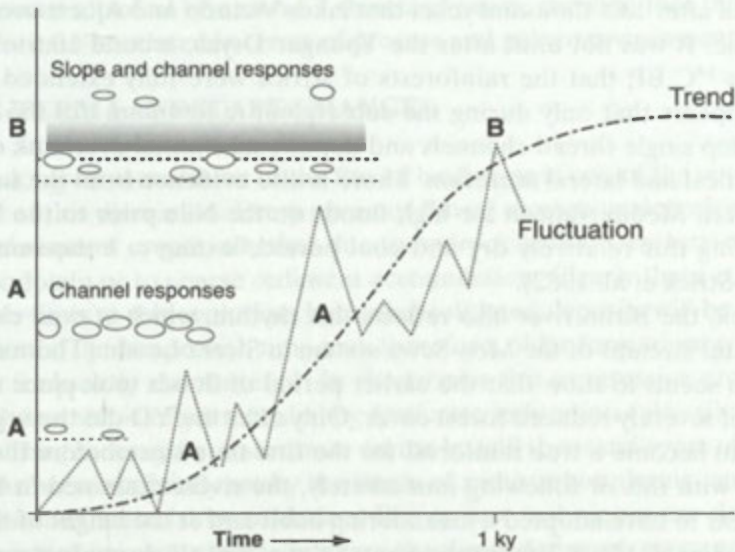


Figure 2. Schematic model to show the geomorphic responses to extreme events during a warming phase (10^2 yr trend curve) of a sub-Milankovich climate cycle. Rainfall is assumed to increase during the period and the impact of short-term fluctuation becomes greater towards the culmination of the warming phase, at B rather than A, although amplitude of the fluctuation may be similar in each case (Reproduced from Thomas 2003, *Sedimentary Geology* 162, *Late Quaternary sediment fluxes from tropical watersheds*, 61-81 with permission from Elsevier Science).

In NE Queensland, it has been shown (Kershaw 1978, Moss and Kershaw 2000) that climate began to deteriorate (becoming cooler, drier) after 79 thousand years, a trend reinforced after 38 thousand years, though earlier clearance by fire used by the recently arrived aboriginal population complicates palaeoclimatic reconstruction (Turney et al. 2001). By c. 26 thousand years open sclerophyll woodland appears to have dominated the area. It is around this time that fan building appears to have begun along the coastal lowlands fronting the east-facing escarpment (Nott et al. 2001, Thomas et al. 2001). At the LGM rainfall may have been reduced by c. 64 percent. Subsequent warming of the post-glacial climate probably began as early as 17-18 thousand years but the rainforest was not fully re-established until after 12 thousand years (10 thousand years ^{14}C BP) (Kershaw 1978, Hopkins et al. 1993). The fans in this area were incised after 15-14 thousand years, and it is thought that this resulted from the kinds of change outlined here.

In Africa, there is ample evidence for an early partial recovery of climate after the cool dry conditions of the LGM (Gasse 2000, Thomas and Thorp 2003), and lakes that had been low or even dry were refilled after 17 thousand years.

By c. 15 thousand years, rivers were depositing coarse sediment in braided channels in West Africa (Thorp and Thomas 1992, Thomas and Thorp 1995), but it was not until after 13.5 thousand years that lakes Victoria and Albert overflowed into the Nile. It was not until after the Younger Dryas, around 10.6 - 9.5 thousand years ^{14}C BP, that the rainforests of Africa were fully extended (Maley 1996). It appears that only during the subsequent millennium did the shallow rivers develop single thread channels and deposit substantial overbank deposits by both vertical and lateral accretion. There is also evidence from the sapropels of the Eastern Mediterranean for high floods on the Nile prior to the YD and also following this relatively dry and cool period, lasting c. 1 thousand years (Rossignol-Strick et al. 1982).

In Ghana, the Birim river also reflects this rhythm, which is even clearer in the headwater streams of the Moa-Sewa system in Sierra Leone (Thomas 1994). The pattern seems to show that the earlier period of floods took place within a landscape of severely reduced forest cover. Only after the YD did the vegetation grow back to become a true rainforest for the first time since before the LGM. Coincident with this or following immediately, the rivers examined in Western Africa appear to have adopted a meandering habit and at the height of the early Holocene 'pluvial' (10-8.2 thousand years) deposited thick muds across their floodplains and low terraces (Thorp and Thomas 1992). The early post-glacial floods brought about scour, undercut banks sending large trees into the river, and deposited gravels and coarse sand in shifting braid bars. The post YD rivers built up floodplains and many tributaries developed swampy valley floors. Subsequent fluctuations of climate led to cut-and-fill episodes, the peaks of successive depositional phases appearing to decline with time. The effect of the YD itself is largely unknown. There are records of low lake levels in Africa and other indications that it was significant in climatic terms. But its impact may well have been only to reduce the overall activity of tropical streams. It has been noted that Vandenberghe (2000) thought the YD might have been too short to bring about significant change to the rivers of Western Europe.

Extreme events will have occurred throughout the climatic oscillations of the late Quaternary, even during periods of reduced rainfall, and cyclone activity. Their significance with respect to long-term landscape development is uncertain. Nott et al. (1996) showed from the study and dating of plunge pool deposits in the Kakadu region of northern Australia that major flood events occurred both during the cool moist climates leading to the LGM, and during the Holocene 'climatic optimum' that followed. But although floods left indelible marks they did not lead to important transitions in fluvial behaviour. In the alluvial fan environment debris flows are likely to be a component of fan building, and can occur within steep mountain streams during intense rainfall. Their effect is often to divert the temporary alluvial channels, but they remain an essential part of the fan environment. In one confined valley of north-eastern Queensland a major boulder deposit, thought to be a debris flow has been buried by 10 m of fine gravels

and sands. The age relationships of these sediments show that the flow deposit is older by several thousand years and that the fine sediment built up by vertical accretion over a period of around 3 thousand years, showing that the contiguous sediments are unrelated in terms of process and palaeoenvironmental setting.

LONG-TERM LANDSCAPE CHANGES

The long-term evolution of the fluvial landscape is one of the more neglected aspects of this discussion. Every phase of fluvial erosion or deposition will influence subsequent events, whether due to bedrock incision, by terracing the alluvial floodplain or to coarse sediment accumulation. Clearly there is no theoretical older limit to this evolution, but unconsolidated deposits will be progressively lost to the landscape with time and, therefore, older formations will in general be less frequently encountered. In the tropics this progressive erosion of past alluvial deposits is often retarded by ferricrete induration (also silcrete and calcrete). The tendency to concentrate on the last 40 thousand years of Quaternary time has been artificially set by the limits of radiocarbon dating, and the loss of carbonaceous material by oxidation. But as new techniques are developed this constraint is removed, especially where the direct dating of quartz grains is possible. What emerges from a fresh look at a number of situations from tropical areas is that landscapes have very different 'age profiles' in respect of their superficial deposits.

In Eastern Zambia, where ancient landslides were found to date before 200 thousand years, samples from local colluvium and alluvium provide evidence for continuous, if episodic, sedimentation throughout the LGC (Thomas and Murray 2001), and similar results have come from neighbouring Tanzania (Sørensen et al. 2001). Evidence for specific and large scale changes at the LGM or since are missing, and the mode of deposition appears to have continued in one section for at least 60 thousand years (OSL). Discontinuous bands of coarse sediment containing cobble-sized clasts are separated by sandy material and the overall form resembles a low angle ($<3^\circ$) fan. Such fans occur widely in the landscape (Thomas 1999), but some sections indicate a single event depositing coarse material in a concave valley head. There are indications that pulses of energy have interrupted the gradual accumulation of fine sediment in a landscape that was possibly initially disturbed from a previous equilibrium at the onset of the LGC (>100 thousand years). This may be important, because the major change in controlling conditions could have come at this time, and subsequent oscillations have been of a lesser order.

In contrast, the Bananal area of E Brazil has provided evidence of major landscape change concentrated around 10 ^{14}C thousand years BP (12 thousand years) (Coelho-Netto, 1997). In Northeastern Queensland, the alluvial fans also appear to date from a specific period between 27/26-15/14 thousand years, after which they were incised (Nott et al. 2001, Thomas et al. 2001). In Sierra Leone

the small rivers appear to retain few deposits prior to 40 thousand years, but the larger rivers and also the Birim in Ghana exhibit widespread high terraces with lateritised alluvium, which remains undated (Thomas 1994, Thomas and Thorp 1995, Thorp and Thomas 1992).

In Western Kalimantan, the major period of sedimentation appears to belong to Isotope Zone 3, which is marked by a consistently developed terrace or series of fans from which finite radiocarbon dates gave ages of 50-55 ^{14}C thousand years (Thorp et al. 1990). These examples suggest that different landscapes have been destabilised at varying times, as a consequence of environmental changes. As yet we do not have a clear explanation for these differences of behaviour, which could be due to regional differences in the nature and severity of climatic and environmental perturbations in the late Quaternary and/or to prior conditions of sensitivity to change.

Bedrock-confined channels are obviously relatively insensitive to climate-induced transformation and will convey water and sediment to lower, unconfined reaches. When overtaken by extreme flood events the high load:water ratio can result in debris flows, which may leave chaotic accumulations of sediment where the channel opens out on to an alluvial plain. The coarse deposits left by these events will subsequently influence the patterns of local sedimentation. But only to the extent that a physical barrier is formed; the behaviour of the river will remain largely a function of load and discharge relationships. Similarly, once a river has become entrenched into former floodplain deposits or coarse fan gravel, the physical energy required to alter the subsequent course and pattern of the river may exceed the magnitude of extreme events, and adjustments, therefore, take place downstream or off-shore. Many rivers exhibit a narrowing of the alluvial plain from older, elevated terraces towards the present-day floodplain. Cut-and-fill structures may become more evident in recent deposits, as the river undergoes internal re-adjustments without the ability to shift course. In many humid tropical fluvial environments, river banks are protected from erosion by riparian vegetation, and the floodplain has formed by vertical accumulation of fine sands and silts, which resist erosion. The progress of vertical accumulation itself ensures that higher and higher floods are necessary to produce over-bank flows. The transition from wide braided courses at the LGM to present-day floodplain morphologies is, therefore, difficult to reverse.

EXTREME EVENTS IN THE SPECTRUM OF QUATERNARY TIME

It can, therefore, be argued that the past history of hillslopes and river valleys leads to real constraints over continuing development, including their responses to extreme events. To some extent this is obvious and should need little emphasis and the interaction of current processes with inherited forms from the Quaternary was emphasised by Starkel in 1987. But the Quaternary timescale is also im-

portant because of the lags and delays in natural systems as they respond to long term climate change, leading us to question the meaning of terms such as 'rapid' or 'abrupt' change. The idea that preparation time is required to condition natural systems to switch between modes of behaviour is not of course new. The concept of slope 'ripening' is used by engineers, and this is related to the concept of 'sediment exhaustion' utilised in geomorphology. We are also used to using the terms 'weathering limited' and 'erosion limited' to describe slope behaviour. And in the tropics, particularly, contrasts between slopes developed on saprolite as opposed to those formed on thin, stony regoliths are great. The amount and degree of weathering achieved in Quaternary time is not fully known, but quoted rates of weathering in crystalline rocks range from 5-50 m Ma⁻¹ so that, over 10⁵-10⁶ yr, many metres of regolith could be produced in humid climates (Thomas et al. 1999).

Another aspect of this discussion is the persistence of sediment stores in the landscape. In the case of widespread colluvium, which is characteristic of many tropical landscapes, this indicates that there has been a lack of hillslope-channel coupling. Large fans and inland terrace units also show that storage of Quaternary sediments is a major feature of many catchments. These sediments are commonly 10⁴ yr in age, even older. This implies that extreme events, which occurred during the Holocene, were unable to remove materials accumulated in the late Pleistocene. In glaciated landscapes, where there are tills and moraines, loess and outwash sediments, it is hardly necessary to emphasise this phenomenon, but in extra-glacial areas the extended duration of uninterrupted fluvial activity reveals more about the nature of long-term landscape evolution.

All the factors discussed here have different but overlapping timescales, which it may be instructive to summarise:

- weathering at rates of 0.01 – 0.05 mm per year,
- vegetation change over 10²-10³ yr,
- changes in magnitude/frequency of extreme events – possibly 10¹-10² yr,
- sub-Milankovich (Bond) cycles c. 1.5 thousand years, embracing climate cooling over 10³ yr; climate warming over 10² yr,
- internal transformation (braiding-meandering, eg.) 10²-10³ yr,
- sediment storage over 10¹-10⁶ yr, implying that sediment exhaustion will be highly variable as a limiting factor in sediment supply.

It is not possible to offer single solutions or models to combine the various factors that influence landscape response to environmental perturbations on different timescales. But it is important to view the occurrence of extreme events that may last hours or days, perhaps weeks, in the context of longer term landscape changes that take place over decades, centuries or millennia. Quaternary spectra are now available from GRIP and GISP2 ice cores and from some ocean sediment cores and these have been used to infer fluctuations in the terrigenous input to the oceans, in some cases apparently matching the incidence of Heinrich and Dansgaard-Oeschger events (Arz et al. 1998). These millennia-scale signals

almost certainly reflect a whole-landscape response to environmental changes including those listed above.

Difficulties are encountered when attempting to construct event-stratigraphies, because sediment delivery to the point of deposition is influenced by a variety of factors (Thomas 2004). Source-to-sink pathways may change within a catchment due to variations in the spatial patterns of rainfall intensity and duration, and this means that different sediment stores may be mobilised during a sequence of rainfall events. To this source of variation has to be added the spatial and temporal aspects of landscape sensitivity to change (Thomas 2001). One aspect of this that has been emphasised here is the complex relationship between vegetation change and sediment yield. Human impacts on plant cover can be immediate and short-term and lead to temporary increases in sediment yield; the way in which long-term vegetation changes operate to increase or decrease landscape sensitivity are less well understood. In the absence of wholesale destruction of natural plant cover, the impact of individual events will depend not only on their inherent characteristics (magnitude, frequency) and antecedent conditions (soil saturation), but also on their position in relation to long-term trends in climate change (Figure 2).

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THE EXTREMENESS OF EXTREME EVENTS

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ABSTRACT: This paper discusses the concept of magnitude and frequency introduced by Wolman and Miller in the middle of the 20th century. The concept is outlined and exemplified from recent examples and reference is made to the need for revision in the light of (a) the interaction between extreme events and human activity; (b) developments in hillslope hydrology, and (c) emergence of our understanding of non-linear behaviour.

The extremeness of extreme events is identified through work-done plots and through conventional statistical probability density functions. It is shown to be controlled (for runoff events) in the short term by vegetation cover, surface crusting and channel network evolution. For the longer term the paper addresses the impact of climatic changes through the vegetation cover by investigating the lagged nature of the response and the amplification or damping of the response through non-linear behaviour.

KEY WORDS: extreme events, logistic behaviour, non-linearity, magnitude and frequency, stability and instability, vegetation.

INTRODUCTION

Extreme events have large impacts on land forms, on human activities and sometimes on national or regional economies. They take the form of intense or sustained rainfalls, droughts or other forms of perturbations to the normal (i.e. modal or average) behaviour of the natural environment. This includes deforestations, dam bursts or intensive soil erosion resulting from meteorological events of high magnitude and relatively rare occurrence. Dry-land environments appear to be especially susceptible to extreme rainfalls, mainly because of the development of convective storm events, sparse vegetation cover and high inter-annual rainfall variability. Because of the economic marginality of dry-land areas, poor people often suffer greatly from extreme events.

MAGNITUDE AND FREQUENCY

A major contribution to the conceptual understanding of extreme events was made as long ago as 1960 by Wolman and Miller in their celebrated paper on the magnitude and frequency of geomorphological processes. This essentially argued that the impact of events resulted from a combination of their magnitude and frequency. Very large events do a lot of geomorphological work but occur infrequently, small events do little work but occur much more frequently (for further discussion see section on Magnitude and frequency revisited). It is the product of magnitude and frequency that results in the most important events, being those of intermediate size and frequency. Wolman and Miller's work was based on river flows and the work they perform. Later workers, among them Starkel (1976) and Froelich and Starkel (1987) and Gallart (1995) showed that the same considerations can be applied to land-sliding and other forms of mass failure. Starkel, Froelich and Soja (1988) showed how both land-sliding and extreme erosion in the Darjeeling Himalayas could be coupled to extreme rainfall events and how they relate to coupling of hillslope and channel systems (Froelich and Starkel 1995).

Richards(1999) has reviewed the magnitude and frequency from the perspective of fluvial geomorphology, and the 'effective event' argument and adduces a number of important criticisms that should be born in mind when applying it. Important as this paper is, the points do not have a strong bearing on the present paper. Indeed they support the main contention here, that extreme event impacts are a function of more than simply extreme rainfall events, and include *inter-alia*, the effects of morphological and climatic pre-conditioning. He also re-iterates Baker's (1977) concern that the relevant stress is the stress-above threshold for a given process and that neglect of this could significantly affect the properties of the functional relationship between stress and discharge.

EXTREME EVENTS AND HUMAN ACTIVITIES

With the preoccupation with global environmental change, interest has turned to the likelihood of greater risks and costs arising from a higher occurrence of extreme events. There has also been a greater recognition of the interaction between extreme events and human activity. Hewitt's (1983) edited contribution on *'Interpretations of Calamity'* considers risks to human communities from spontaneously occurring geophysical events such as storms, floods, and unseasonable cold or drought. A study of the Vaison-La Romaine extreme rainfall in the Ouveze catchment (SE France) by Wainwright (1996) demonstrated the importance of hillslope cultivation and the modification of river channel beds by human activities for extreme runoff generation.

In the volume edited by Starkel and Basu (2002), Starkel discusses the role of extreme events in the context of environmental changes in the mountains of the monsoonal tropics. It would be difficult to imagine more difficult environments for which to make a contribution to management.

HILLSLOPE HYDROLOGY AND NON-LINEARITY

Two other contributions have changed our perception of extreme events. First was the development of hillslope hydrology initiated by Hewlett and Hibbert (1967) and developed by Kirkby and Chorley (1967) and Kirkby (1969 and 1978). Secondly, there was the recognition of the non-linear behaviour of geomorphological systems by Schumm (1973) and Brunnsden and Thornes (1979). The first contribution highlighted the importance of intrinsic thresholds and their role in determining the flood and sediment hydrology of river basins. Flood events could no longer be understood in terms of the simple rainfall-runoff models of 1960's hydrology. The second contribution led to the structuring of rainfall-runoff relations in terms of thresholds and event reaction and relaxation times. It introduced the idea of the 'sensitivity' of landforms in terms of the relaxation times after events of high magnitude, with sensitive landscapes being those with mean event spacing being shorter than the mean relaxation times (the concept of transient form ratios). Increasingly, it has been demonstrated that non-linear behaviour of geomorphological process is important in the process-response relationships that underlie geomorphological understanding (Thornes 2003).

Until recently, papers on extreme events have largely comprised descriptions of the events and their consequences, and have rarely entered into debate of a conceptual or theoretical nature. It has thus been difficult to draw general conclusions or create general theory about extreme events. This chapter attempts to focus on two new aspects. First the notion of the extremeness of extreme events and what controls them. Second, with particular reference to semi-arid environments, the role of vegetation cover and how it determines the persistence of the impact of an extreme event on hillslope runoff and sediment systems in time.

THE EXTREMENESS OF EXTREME EVENTS

It has already been noted, following Wainwright (1996), that human activity increases the impact of extreme events by producing surfaces that generate more runoff, and that reduce the time taken for runoff to reach the channel. These effects have been explored by hydrologists for over a century. Four other themes (that relate to the extremeness of extreme events) have emerged much more recently. They are: (a) how the magnitude and frequency of different processes change at different time scales, (b) the importance of the different hydrological responses of different soil surfaces in a catchment, (c) the role played by hillslope-channel coupling and (d) the self-organisation of channel networks within catchments, which results in greater efficiency of water and sediment transfer to the main channel, which in turn tends to increase the extremeness of events.

MAGNITUDE AND FREQUENCY REVISITED

As Gallart (1995) points out, the Wolman and Miller (1960) magnitude and

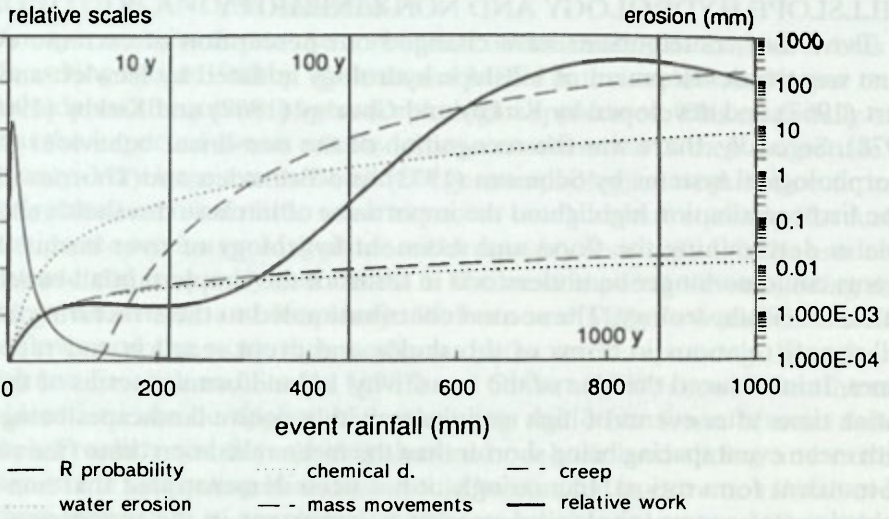


Figure 1. Work done by four processes at different rainfall event in the Llobregat catchment, Spanish Pyrenees. From F. Gallart, 1995, p.359.

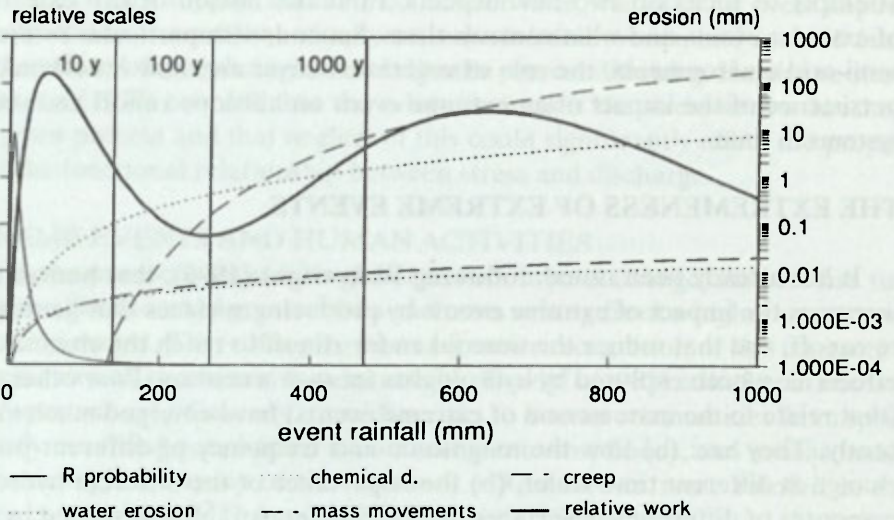


Figure 2. As figure 1. The same four processes are recorded, but the rainfall frequency distribution used to calculate the work done by each process (and then summed to provide the total work done) has a lower standard deviation (smaller spread) than that used for the calculations of the result shown in figure 1. As a result, the pattern of total work done by events of different magnitudes and frequency is also changed. From F. Gallart, 1995 p.361.

frequency proposition arises mainly because the product of the power function (work done by flow) and the negative frequency exponential (the frequency distribution curve of events causing the flow) tend towards zero for very high values. The difference between the most frequent (modal) events and those which do the most work depends on the exponent of the power function and the standard deviation of the frequency distribution. Other kinds of functions for the work-done curve can produce a maximum for 'trigger' values whose frequency is very low. He demonstrates that (as identified by Starkel 1976 for the Darjeeling Himalayas) the most important processes in mountains are rapid mass movements, the number of which increases very quickly with increasing rainfall. This is demonstrated by plots of the number of landslides against the magnitude of the events which produced them. Although the slopes of the assembled curves are similar for one-day events, the intercepts are quite different, showing that the same rainfall amount produces drastically different landslide occurrence depending on the climatic and geomorphological scenarios.

By investigating the magnitude-frequency distributions for four processes in the Llobregat basin (Spanish Pyrenees), Gallart shows that, at frequencies of less than the 100 year event, chemical denudation, creep and wash are most important but that the total work-done curve is dominated by mass movements generated by rainfalls of events of c. 1000 year or more recurrence (Figure 1). By repeating the analysis with a less variable rainfall, the relative roles of the different processes vary and the work-done curve is sharply bimodal in form (Figure 2). The first mode at the c.10 year event, is a combination of the three most frequent processes and the second mode, the mass movement mode, is lower in magnitude and shifted towards less frequent events (c. 600 yrs) These results were anticipated, but not empirically demonstrated by Carson and Kirkby (1972).

HYDROLOGICAL RESPONSE SURFACES

It is well established that the vegetation cover plays an important role in determining runoff and sediment yield magnitudes for rainfall events. In addition it has been adopted as almost axiomatic that forest cover is the optimal approach to reducing runoff and sediment yield. In fact Nortcliff and Thornes (1977) demonstrated in Amazonas that, in extreme events, the canopy fills so quickly that runoff is not proportionately reduced. In a similar fashion the widespread assumption that transmission losses significantly reduce the peakedness of large flows in ephemeral channels was not found to be justified by the results of digital modelling of Butcher and Thornes (1979).

As well as the effects of vegetation cover, other surface textures play a significant role in increasing or reducing the extremeness of extreme events, for example through crusting and sealing. There is a great amount of evidence that crusting plays a central role at the level of the crop water balance, as well at coarser scales of analysis (Patrick 2000). In semi-arid areas, runoff potential for water harvesting can be enhanced through modification of surface crusting processes,

which are widely occurring phenomena. Patrick (2000) found that crust morphology was a function of environmental variables controlling crust genesis and that a crust's appearance in the field could therefore be linked to infiltration values. Escadafel (1989) has developed a methodology for mapping infiltration from remote sensing, establishing the potential for integrating hydrological response surfaces with hillslope models.

TOPOGRAPHIC CONTROLS IN HILLSLOPE-CHANNEL COUPLING

Kirkby, Bracken and Reaney (2002) studied the influence of land use, soils and topography on the delivery of storm hillslope runoff to channels in the Nogalte river basin in Murcia, southeast Spain. The runoff is 'excess' runoff, i.e. the amount of runoff produced after a runoff threshold has been exceeded. The main controls on the threshold rainfall are interception by plants leaves and stems, interception on litter, depression storage and storage within the soil. These are co-linear with and related to slope and prior soil moisture conditions. Depression storage is strongly related to gradient. Topography at different scales is shown (by digital simulation) to play an important role in delivery at different scales. The ploughing direction (angle of furrow relative to contours) also plays an important role in depression storage and hence runoff (Figure 3). Furrows that are parallel to the contour have a far greater depression storage than a purely random roughness pattern. At any other furrow orientation, depression storage is always less than the purely random pattern.

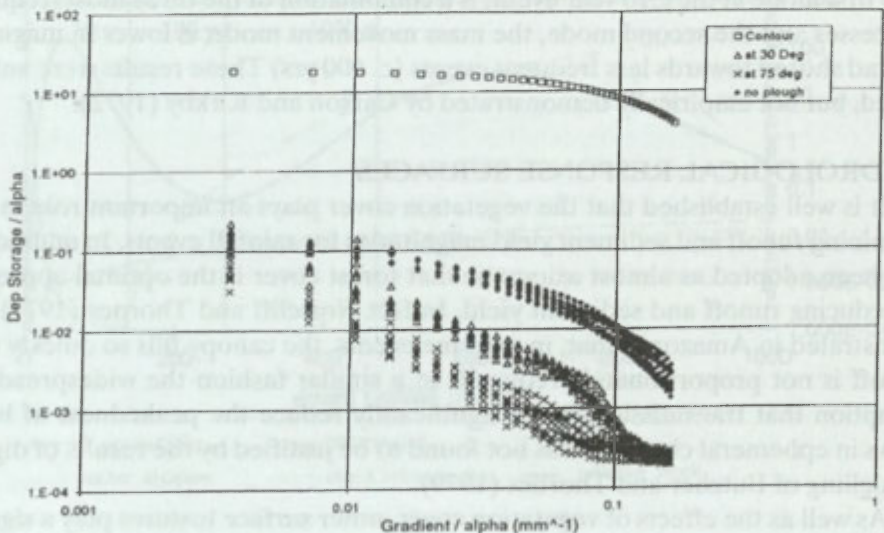


Figure 3. Depression storage associated with random roughness imposed on regular furrows, inclined at different angles to the contour. The unploughed case is also included for comparison.

From Kirkby, Bracken and Reaney, 2002.

THE CHANNEL NETWORK

Drainage density plays a critical role in the extremeness of runoff as indicated by multivariate studies in the last century. Rodriguez Iturbe (1993) has demonstrated analytically that channel network characteristics tend to evolve through feedback (by network extension) to adapt to the prevailing hydroclimatic environment in which the system is operating.

These four topics illustrate that it is important to develop a deeper understanding of the mechanisms by which extreme rainfall events produce different geomorphological outcomes if reasonable management strategies are to develop beyond the afforestation palliative that is so widely advocated.

IN THE LONGER TERM

Geomorphologists in the middle of the last century frequently attempted to interpret earth history from sedimentological and geomorphological (form) evidence. In this context extreme events were special because the rate of change in them was very fast. Geomorphological evidence (old soil erosion surfaces) could represent centuries, or decades of erosion or even a single huge storm. For the Holocene this led to real difficulties in interpreting the impacts of human activities. Moreover, a pre-requisite to interpretation was a contemporaneity between form and the event that produced it. If there was a strong lag between an event and the effect it produced, then palaeo-reconstruction was made even more difficult. In addition the cause-effect was often assumed to be linear—a big event produces a big effect and vice-versa, hence the special interest in magnitude and frequency.

There are two questions to be addressed in relation to extreme events in this context. First, if there is a step change in climate, how long would this take to reveal itself in morphological changes. Second, could large responses arise from very small climatic changes, climate variations or even single extreme events? These two questions are addressed in this section, through the vegetation cover in dry environments which appear especially sensitive to variations, fluctuations and extreme events.

THEORETICAL CONSIDERATIONS

Since May's 1973 paper, population biologists, geneticists and evolutionary ecologists have recognized the effects and problems associated with the impact of a probabilistically varying environment on biological population dynamics. May investigated discrete growth under density-dependent conditions, showing that the populations could exhibit pathological behaviour in which they might be stable, might exhibit stable cyclical behaviour or deterministic chaos. This had the important consequence that stable cyclical behaviour of populations could arise intrinsically, without the need for strongly cyclical external forcing (usually

assumed to be climatic). The consequences for unstable vegetation systems and for erosion have been demonstrated for semi-arid environments (1988 and 1990; Thornes and Brandt 1993). This non-linear behaviour also has deep ramifications for the understanding of global environmental change (Thornes 2003).

TIME VARYING ENVIRONMENTS

In the same paper, May sought to evaluate the effects of time varying environments through the logistic differential equation that has a long and distinguished history in biology and is reproduced as equation 1 below. This approach has been used (Thornes 2003) to investigate the impacts of climatic variations on the bush-grass boundary in South Africa. If we assume that soil conditions remain constant, how do variations in the annual rainfall and the potential vegetation growth affect the long term direction of the vegetation cover? Put slightly differently, how long does it take the effects of extreme wet or dry years to be felt in the vegetation cover? And how likely is it that the cover will be 'tripped' towards complete cover or towards bare soil with the well-documented impacts on soil erosion (none or catastrophic)? What characteristics of the environmental fluctuations are critical in propagating the effects of change through the eco-hydrological system?

May shows that 'There is no difficulty in solving the logistic differential equation for arbitrarily time-dependent $r(t)$ and $K(t)$ ' (p.20). (Note that r there is the K in equation 1 below, and $K(t)$ is the $V_{cap}(t)$ in equation 1 below). When r (May's notation) is constant, $N(t)$ tends, after sufficiently long time, to an asymptomatic value. In the simulation study described below the 'population' is $V(t)$, the above-ground biomass. The population's characteristic response time is $TR=1/r$ (May, page 21) and $N(t)$ is the inverse of some weighted harmonic average over past values of the carrying capacity.

If the carrying capacity, V_{cap} , varies periodically, the behaviour of the population $N(t)$ will now depend on whether its characteristic response time is large or short compared to the period (τ) of the environmental oscillations. In the limit when TR greatly exceeds τ the population averages out the environmental variations at less than the average of $K(t)$. When TR is short compared with τ , the population tends to 'track' the environmental variations. The simulation results obtained below conform to these theoretical results reported by May.

In evolution, genes are continually subject to fluctuations in their (chemical) environment and Roughgarden (1979) has theoretically analysed and modelled how these stochastic fluctuations affect the genome make-up. This too is built on the initial work of May (1973) on demographic stochasticity.

Roughgarden points out that most models are based on the assumptions that the parameters in the model are constants, but that the carrying capacity varies. He also adopts two kinds of stochasticity, demographic (intrinsic) and environmental (extrinsic). One of the most important implications of environmental

fluctuations for this paper is that population extinction can result (vegetation cover can drop to zero).

The probability of various population sizes at time t is described by Roughgarden (1979) using the diffusion equation that, as usual, comprises two parts, a deterministic part and a stochastic part. For genetic drift, the stochasticity arises from independent sampling of the gene pool at each generation. In fact there is a problem because diffusion theory assumes independent fluctuations in time, whereas climate series are almost invariably auto-correlated, but the diffusion approximation is robust enough to tolerate this departure (Roughgarden, p. 377).

Roughgarden discusses two methods of solving the diffusion equation (the Ito method and the Stratanovich method) and from this discusses the attractors. These are the boundaries to the system. In our case they are (i) complete vegetation cover, (ii) 30 percent vegetation cover – recoverable catastrophic erosion and (iii) 0 percent vegetation cover – irreversible catastrophic erosion. The 30 percent value is based on the proposition that at or above this value of cover, erosion is practically eliminated (Thornes and Francis 1990).

This digression into population biology demonstrates that there is a theoretical basis to the question: How big (how extreme) do perturbations have to be to shift the eco-hydrological system to an irreversible situation and how long will it take? A boundary (attractor) is said to be unattainable if sample paths do not reach the boundary in finite time. In the following section it is considered how different characteristics of V_{cap} and the growth coefficient (K) through time determine how close the respective boundaries are after 100 years and how quickly specific values of the vegetation cover are reached after a perturbation to the starting values of cover under relatively wet and relatively dry rainfall regimes. Fluctuations in V_{cap} (the rainfall limited potential above ground biomass) and K (the logistic growth coefficient) are considered.

SIMULATIONS

Another way to examine the effect of rainfall events on plant-growth, and hence on runoff and erosion, is to simulate the behaviour by computer modelling (Mulligan 1998, Thornes 1990). Mulligan simulated the geomorphological impacts of climatic variability and questioned the assumption that environments trend towards equilibrium and are only displaced by strong external forcing. He concluded (p. 60) that:

‘In Mediterranean environments, the geomorphic significance of climate variability may be as great as that of climate change, particularly over short time scales.’

He also demonstrated the importance of functional vegetational types in determining the impact of plant cover on extreme events and geomorphological responses.

Thornes (1978 and 1990) adopted the logistic equation as the basis for plant

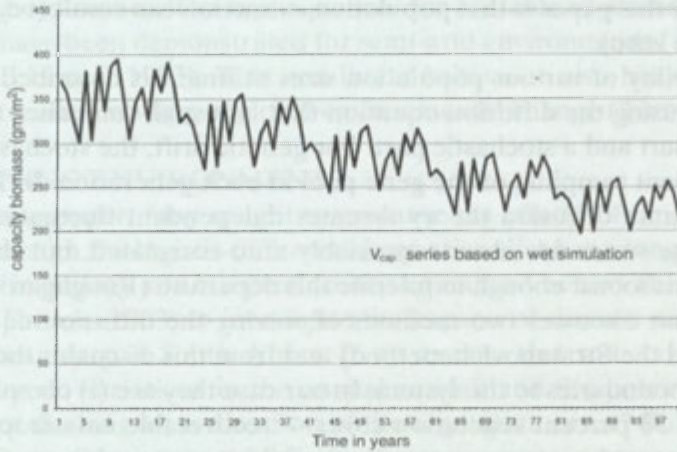


Figure 4a. V_{cap} series for 100 years, based on rainfall showing regular oscillations typical of ENSO affected rainfall regimes. An overall decline has been superimposed using a negative autoregressive correlation after first 20 years. The initial value is 370gm/m^2 .

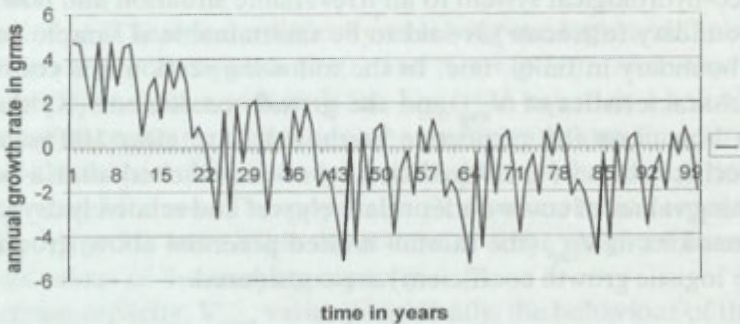


Figure 4b. Actual growth rate (gm/m^2) according to logistic equation when forced with the series shown in Figure 4a and with $K=0.06$.

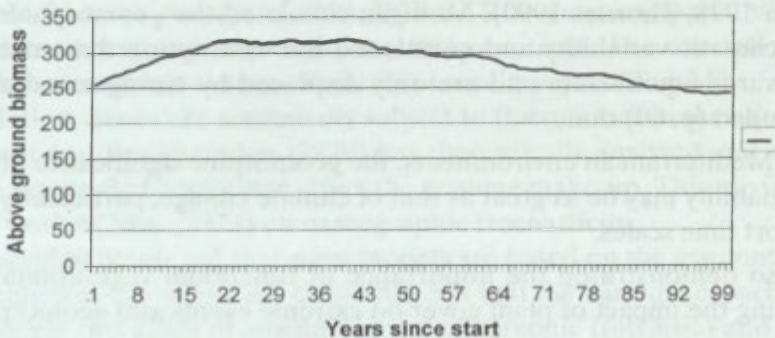


Figure 4c. Simulated vegetation biomass (gm/m^2) for wet series using logistic equation and with $K=0.06$.

growth simulation. This is:

$$dv/dt = K \cdot V_{(t-1)} [1 - V_{(t-1)} / V_{cap}] \tag{1}$$

$$\text{and } V_t = V_{(t-1)} + dv/dt \tag{2}$$

In order to be able to generate the growth according to the logistic equation (equation 1) the value of V_{cap} has to be supplied. This is the maximum possible biomass (gm/m^2) that can grow above the ground according to the available soil moisture. The latter is a function of rainfall and temperature in dry environment, as well as soil characteristics. In this analysis it is assumed that soil is uniform and that only rainfall is varying to produce V_{cap} . Given V_{cap} the logistic equation can be used to obtain the vegetation growth rate, that is then added to the V of the previous year (equation 2). Figure 4a shows the V_{cap} series based on a starting value of $350 gm/m^2$. The first 20 years are supplied from an actual rainfall series in the Eastern Cape, South Africa (Thornes, 2003), but comparable values occur in S.E. Spain. The graph shows a strong oscillation reflecting the El Nino Southern Oscillation (ENSO) events and an overall decline in capacity induced artificially by an autoregressive process. After 100 years. the V_{cap} has fallen to a low of $200 gm/m^2$.

Figure 4b shows the annual growth rate in $gms/sq m/yr$ as determined by the logistic equation when the value of K is set at 0.06 . Note that the growth rate can be negative as well as positive and goes negative after about 22 years and remains dominantly negative for the remainder of the 100 years. This is to be expected, given the steady decline of V_{cap} due to the modelled falling rain.

Figure 4c shows how the above-ground biomass is smoothed compared with the rainfall fluctuations by the logistic behaviour. This is a damping of the strongly oscillatory behaviour of earlier models (Thornes and Brandt 1993). If the rainfall is exceptionally high, the overshoot of V_{cap} is compensated for by a strongly negative growth. Respiratory losses in biomass are not modelled here (c.f. Thornes, 1990).

Figures 5a and b illustrate the effects of varying the logistic K value and can be compared with Figure 4c. Damping is greatest when $K = 0.01$, rises to $K = 0.05$ and falls again to $K = 0.09$. These results are also reflected by the summary statistics given in Table 1. Here the variance in the biomass series rises and falls in the same manner. Another result of comparable interest is the time taken to reach a given value of biomass.

Similar results are obtained for the 'dry' series. In Figure 6a the dry series of V_{cap} is observed to oscillate strongly, but the logistic series (6b), even with $K = 0.06$, is substantially damped.

As in the last section, the question at issue is when does the vegetation first reach a given value (e.g., $250gms/m^2$ or $0 gm/m^2$) or what value is reached after a given time (e.g., 100 years). As before, these are expected to depend on (i) the initial value (V_0), (ii) the K value in the logistic equation and (iii) the variation in the V_{cap} values (the driving variable). For the dry series the variation in first

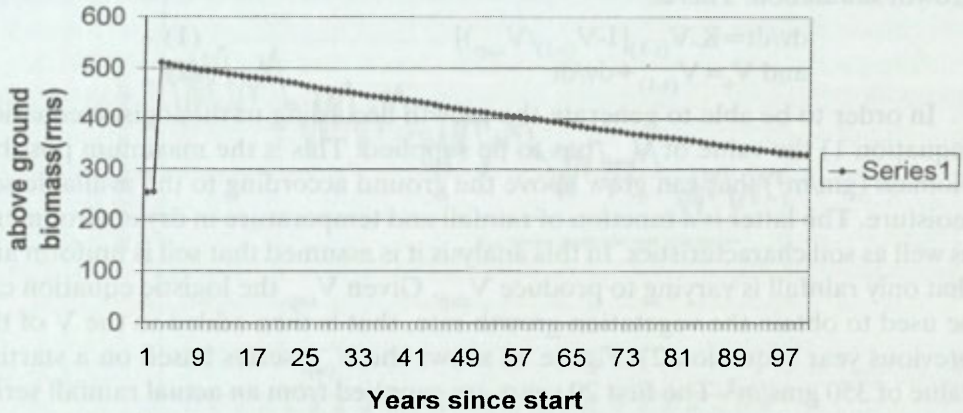


Figure 5a. Simulated vegetation biomass (gm/m^2) for wet series according to logistic equation when forced with the series shown in Figure 4a and with $K=0.01$.

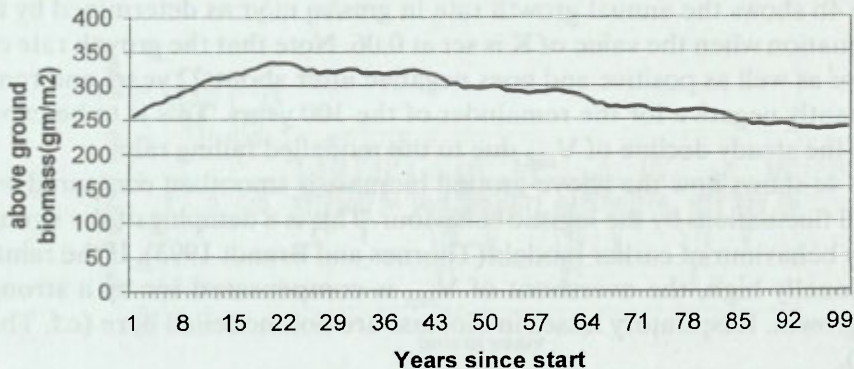


Figure 5b. Simulated vegetation biomass (gm/m^2) for wet series according to logistic equation when forced with the series shown in Figure 4a and with $K=0.09$.

time to 250gm/m^2 and the V_{100} are given in Table 1. It shows that, under the simulated conditions, $v = 250\text{gm/m}^2$ is reached after 30 years for $K = 0.01$ and after 13 years when $K = 0.03$. When $K = 0.05$, the value is reached almost instantaneously and thereafter the value is reached with the first adjustment because the starting value is very close to 250 grams. Although the system has not been run to zero, the value reached after 100 years does show that, as K increases, the rate of fall is significantly affected by K , such that the value of V_{100} is less as K approaches 0.05, after which it stabilises notably to around 160 gm/m^2 .

The table shows the statistical characteristic of the key variables (V_{cap} and V , both grms/m^2) simulated under dry and wet rainfall conditions. The mean value of V_{cap} for the dry series is 202.5 and for the wet series is 286.7. For the dry series,

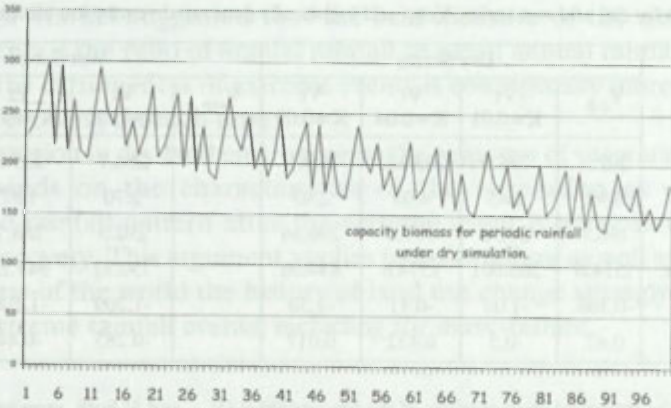


Figure 6a. V_{cap} values for dry series. An overall decline has been superimposed using a negative autoregressive correlation after first 20 years. The initial value is 230gm/m^2 .

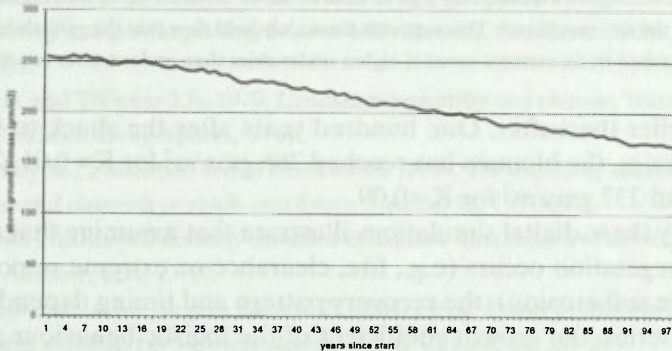


Figure 6b. Simulated vegetation biomass (gm/m^2) for dry series according to logistic equation when forced with the series shown in Figure 6a and with $K=0.06$.

the variability of V_{cap} is almost twice that for the wet series. This reflects the widely observed higher variability of rainfall in dry conditions than in wetter conditions.

The table also illustrates how the characteristics of the vegetation biomass ($V_t, \text{gm/m}^2$) vary according to the K parameter of the logistic equation for a given rainfall series. It varies more with the higher K in the dry conditions than in the wet conditions. This supports the widely held view that the probability of a critical threshold being reached by an extreme event is higher under drier than under wetter conditions.

The wet series starts at 370gms/m^2 and therefore (as shown in figure 4b) the growth is positive in the early years, but later becomes negative. Table 1 shows that the peak biomass is delayed systematically as K increases to 0.05 and the peak

Table 1. Statistics for dry and wet series, V_{cap} and logistically generated vegetation biomass.

	Dry Series				Wet Series			
	V _{cap}	V _t K=0.01	V _t K=0.04	V _t K=0.09	V _{cap}	V _t K=0.01	V _t K=0.04	V _t K=0.09
Mean	202.5	238.0	203.123	209.1		286.7	285.0	287.4
Standard error	3.93	1.45	4.03	2.93		2.70	1.87	2.86
Median	196.7	241.2	197.4	206.36		292.5	288.7	291.8
Sample variance	1514.6	205.001	1574.6	834.96		730.32	347.22	914.5
Kurtosis	-0.366	-1.06	-0.41	-1.30		-1.209	-1.108	-1.204
Skewness	0.42	-0.5	0.432	0.017		-0.285	-0.44	-0.25

The table shows the statistical characteristic of the key variables (V_{cap} and V , both grms/m^2) simulated under dry and wet rainfall conditions. The mean value of V_{cap} for the dry series is 202.5 and for the wet series is 286.7. For the dry series, the variability of V_{cap} is almost twice that for the wet series. This reflects the widely observed higher variability of rainfall in dry conditions than in wetter conditions. The table also illustrates how the characteristics of the vegetation biomass (V_t , gm/m^2) vary according to the k parameter of the logistic equation for a given rainfall series. It varies more with higher K in the dry conditions than in the wet conditions. This supports the widely held view that the probability of a critical threshold being reached by an extreme event is higher under drier than under wetter conditions.

is reached earlier thereafter. One hundred years after the shock to the system, with the wet series, the biomass has reached 265 gms/m^2 for $K=0.01$, 245 gms/m^2 for $K=0.05$ and 237 gms/m^2 for $K=0.09$.

In summary these digital simulations illustrate that assuming that an extreme shock to the vegetation occurs (e.g., fire, clearance or extreme removal by grazing or massive soil erosion), the recovery pattern and timing depend on the ensuing rainfall series, the growth coefficient of the logistic behaviour and the variability of the driving force (rainfall in this case), in compliance with the theoretical behaviour described above. Moreover, if K can be thought of as a water-use efficiency coefficient, then over long periods K itself could evolve in such a way as to adapt the vegetation to the prevailing environmental conditions (as suggested by Eagleson (1982)). Moreover, different species with different K values can be expected to recover from extreme events at different rates, a factor which needs to be taken into account when deciding on a mitigation strategy for land degradation.

CONCLUSIONS

The role of extreme events is important for a variety of geomorphological processes and has to be accommodated in problems of environmental management as firmly established by Starkel (1976) and Starkel and Froelich (2000) for the Darjeeling Himalayas and by others for dry-lands in the Mediterranean, the United States, Canada and Southern Africa.

Although Starkel suggested that the best measure of the extremeness of extreme events is the ratio of annual rainfall to mean annual rainfall, it is shown above that the extremeness of extreme events is conceptually more complicated and is itself evolving through time.

For soil erosion in dry-land environments the response of vegetation to extreme events depends on the characteristics of the vegetation as well as later rainfall. The rainfall pattern after the extreme event is particularly important for system recovery. This argument applies to mass-failure as well as soil erosion. In many parts of the world the history of land use change strongly modifies the impact of extreme rainfall events, including for mass-failure.

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DIVERSITY AND MAJOR CHARACTERISTICS OF NATURAL AND DISASTERS

Natural disasters are of wide-spread occurrence throughout the world. They vary in type, intensity and frequency with the environmental situation and their disastrous effects on society depend on factors such as land utilization, population

NATURAL AND HUMAN FACTORS IN ENVIRONMENTAL DISASTERS

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ABSTRACT: The world is approaching limits of growth and consequently facing depletion of natural resources and widespread environmental degradation. The sustainability of many present economic activities is problematic and the interaction between humanity and environment more crucial than ever. The critical situation is recognized in scientific and engineering circles and by decision makers at all levels. Creeping hazards and related disasters, such as land degradation and desertification, are on the rise and instantaneous environmental disasters are of growing concern too. Natural hazards of exogenous origin, such as floods and landslides, have natural as well as human causes while those of endogenous origin, such as volcanic eruptions and earthquakes, have natural causes only. However, the disasters associated with hazards of any kind are particularly severe in densely populated regions and in areas more vulnerable to extreme events because of environmental and/or economic marginality. Extreme events are most effective as a destructive element where environmental degradation has been provoked by inappropriate land utilization. Climatic changes and lack of awareness and preparedness are aggravating factors. Modern scientific concepts and emerging powerful technologies provide new tools for addressing the problem of balancing human needs and environmental equity.

KEY WORDS: aerospace technology, creeping disasters, disaster mitigation, early warning, extreme events, hazard zoning, natural disasters, sustainability, vulnerability,

DIVERSITY AND MAJOR CHARACTERISTICS OF HAZARDS AND DISASTERS

Natural hazards are of wide-spread occurrence throughout the world. They vary in type, intensity and frequency with the environmental situation and their disastrous effects on society depend on factors such as land utilization, popula-

tion density and coping strategies. Natural disasters are, in fact, not always quite natural but can be triggered or aggravated by human activities, such as inappropriate land utilization or engineering works. Society may be a causative factor as well as a victim of disasters. The two main groups of disasters are creeping disasters and instantaneous disasters. Creeping hazards and disasters, such as land degradation and desertification, are primarily linked to environmental stress caused by changes in land use and/or climate. Disasters of the instantaneous group relate to, often spectacular, extreme events in the first place and result in overnight changes of the land.

Creeping hazard assessment is rooted in environmental research and in social science as well. The sustainable use of the resources of our planet is essential for safeguarding our life support systems on the global, regional and local levels. It requires careful matching with the utilization and regeneration of resources by keeping economic efficiency, environmental equity and social justice in balance. That economic activities may cause ecological stress is well-known but that social conditions are connected with unsustainability too is often underrated or ignored, however. The problem is not new: for example deforestation and land degradation also existed in the Greek and Roman empires in the Mediterranean area, in ancient China, in Latin America, etc. and have since long affected even sparsely populated savanna areas where annual burning of grass cover was (and is) common practice. There are many cases on record where the situation went out of control, and ultimately only denuded and deeply eroded slopes and infertile badlands remained. In some cases a new and entirely different equilibrium subsequently developed in the impoverished environment. The present situation is particularly disquieting, however, because of the universality of rapidly increasing economic activities and the sharply rising world population. There are also many examples in the past, however, where ingenious methods of land management developed that ensured sustainability. A great diversity of such ancient technologies has been discovered and traces can often still be found in the form of terraced slopes, irrigation systems, etc. They give us some hope that with modern science and technology, it should be possible to find proper solutions for the dramatic global sustainability problems of our time.

Instantaneous disasters can be divided in two main categories: those of endogenous origin and those of exogenous origin. The first one includes extreme events such as earthquakes, volcanic eruptions, tsunamis, etc. that are completely beyond our control and of which the moment, location and magnitude are often hardly predictable, particularly in the case of earthquakes. They may cause drastic landform changes, devastating society in the process. Hazard zoning, rooted mainly in the geomorphological lay of the land, is an important means of reducing loss of human life and property. It only is effective, however, where related physical planning is a real option. Bad governance, scarcity of safe ground, economic importance of endangered areas, financial problems, social institutions, etc. limit optimal use of hazard zoning in many parts of the world. Early warning

systems (Scott et al. 1997), if applicable, are another means of disaster reduction, provided that an emergency scenario is at the disposal of competent authorities and that the population is aware of the hazard and prepared for a disaster. The second category, disasters of exogenous origin, includes landslides and other mass movements, river floods, coastal flooding by sea surges or by strong winds or cyclones. The causes of these events are natural in the first place, but the effects of human impact on the environment are becoming increasingly important. The present, at least partly man-induced, climatic changes are now recognized as a new major element. The reduction of these kinds of disasters is rooted in environmental management and in hazard zoning and early warnings as well.

Global non-renewable resources, such as iron, coal, oil and gas, are being depleted at an alarming rate. Recycling of materials such as iron only stretch their life span, and introduction of other, synthetic materials to replace them is essential for sustainability. The global energy problem is another major issue. Good management of the global renewable resources, such as soil, water and forest, is of particular importance in the context of natural hazards and disasters. They are under severe stress and this easily results in soil erosion, land degradation, river floods, drought and desertification. Pollution of soil, water and air, problems of urban waste disposal and animal manure are other signs that there is something basically wrong with many modern production systems and with our consumption habits: there is no sustainability as a result of the steadily rising population and per capita consumption that lead to a hardly controllable drive for economic growth with disregard of environmental carrying capacity. Introduction of innovating technologies and new scientific approaches, coupling high productivity and conservation practices, is required in combination with lifestyle changes.

Achieving sustainability – and thus reducing environmental hazards and disasters – is complicated by the fact that we are in a dynamic situation. The world is continuously changing in various respects and the rates of change are rising dramatically. These changes comprise environmental changes as well as changes of society. Environmental changes relate in part to natural causes that are beyond our control. However, they are also, at least in part, triggered by human interference with vegetation cover, composition of the atmosphere, etc. The rapidly rising average air temperatures recorded all over the world are an important element of global environmental change, the causes of which are at least partly situated in human action. It is, with justification, a focus of international research. The distributional patterns of floods and droughts are changing and their frequencies and intensities increasing. Rising sea levels, caused by global warming, lead to an increase of coastal flood disasters. It is evident that many kinds of hazards and disasters are related to environmental degradation caused by unsustainable human activities. This fact is clearly understood by the scientific community at large and was formulated by W. Steffen of the International Geosphere Biosphere Program (IGBP) as follows:

'We are on a turning point in the evolution of global science. Humans are now recognized as a critical element of the earthsystem itself, as an interacting agent of change as well as a responsive recipient of the impacts of changes in the system. Research that ignores the human dimension is becoming increasingly irrelevant.'

EXTREME EVENTS, CREEPING DISASTERS AND SUSTAINABILITY

The disastrous effects of extreme events on society include death toll, injured people, uprooted population, capital losses and also impeded socio-economic development as a result of destroyed infra-structures and production units and of social or family structures. A number of major natural disasters that occurred during the 1980s made people throughout the world aware of the disaster enigma and prompted the United Nations to launch the International Decade for Natural Disaster Reduction (IDNDR) that came to an end in the year 2000 and subsequently was followed by the UN International Strategy for Disaster Reduction (ISDR) that is now well underway. Because the roots of natural disasters are not always natural and technical-industrial disasters are increasingly important too, the word 'natural' has been omitted in the name of this new international program. It is a response to the rising frequency and the rapidly increasing impact of extreme events on society.

More people are at present adversely affected and economic losses are higher than ever, notwithstanding the efforts and successes of the IDNDR. About 25 percent of the population of the world is at risk and the number of casualties varies between 50,000 in quiet years and 300,000 or more during one major flood or quake event. The economic losses have more than tripled in the last 30 years and are now in the order €100 billion annually. The insurance bill is around €15 billion Euro 'only' because of low insurance density in many affected areas, particularly in developing countries. Windstorms and floods account for about 85 percent of the economic losses and 90 percent of the insured losses. Insurance companies re-ensure themselves against the potentially massive claims from their clients. Occurrence, location, magnitude and frequency of natural disasters are determined by two main elements: the susceptibility of the land to certain, creeping or high intensity - low frequency, events on one hand and the vulnerability of the affected society on the other. Because these elements are interactive, disaster issues are at present placed in the context of environment, life support systems and sustainable development (Zschau J. and Küppers 2003).

Natural climatic fluctuations, covering periods of several years or decades, are a very distressing element in hazard assessment, complicating efforts of combating flood-and-drought disasters. It is essentially distinct from continual environmental change and is nowadays often referred to as the El Niño effect. This term relates primarily to oceanic and atmospheric fluctuations in the equatorial

Pacific region but is linked to other, comparable phenomena elsewhere in the world, such as the Southern Oscillation in the Indian Ocean and SE Asia, the North Atlantic Oscillation, etc. It is, in fact, a global issue, that, for instance, recently caused extensive bush fires, air pollution and health problems in Indonesia, Australia and other countries, where it also adversely affected food production. The droughts, still fresh in memory, that some decades ago have ravaged large parts of sub-Saharan Africa, were also caused by these oscillations that furthermore led to extreme river floods in other regions and affected the tracks of tropical cyclones. The extreme and repeated floods of some European rivers, such as the Meuse, Rhine and Oder, in the last decade (German National IHP Committee 1997) are thought to result from global warming, a supposedly man-induced continuous process of change. Whether climatic fluctuations play part in it too, is unclear. Rising population densities evidently increase vulnerability and risk. There are, however, in every society that is vulnerable to drought diverse traditional coping strategies, that range from water harvesting techniques to temporary changes in social organization. These strategies merit our full attention and research: they provide an insight into the environmental situation and can assist in finding modern solutions for our present problems. The same applies to traditional indigenous coping strategies with floods and other types of environmental crises that exist in endangered societies.

Changes in society are another important field of change that is on par with changes of the environment in the context of sustainability problems. A major issue is the increasing population density that gives rise to growing pressure on the land in rural areas and to rapid urban growth. The high and still sharply rising per capita consumption in the industrialized world is another main factor of environmental stress endangering sustainability. But the desire for a better living is, of course, universal. Today's have-nots want to get a better share in the future. Global rise of the production/consumption by and the quality of life of the have-nots is a must for global stability, and the solution of the related global economic, social and environmental issues should get top priority. The present contrasts are enormous and certainly not sustainable. For example: 25 percent of the GNP of the African continent emanates from within 100 km of Johannesburg. Social changes leading to depopulation of marginal rural areas may result not only in a decreased cohesion of rural societies but also in land degradation due to lack of farmhands to maintain conservation and irrigation works. The introduction of alien tenure systems and monocultivation may also conflict with proper management of land resources. Clear land ownership regulations are an essential land management tool, notably in densely populated rural areas where steep slopes are deforested and put under agricultural use. The farmers will be prepared to do the laborious conservation work, terracing, etc. only if they know that the land is irrevocably theirs. The village communities provide a structural basis for implementing and maintaining the local environmental resources.

The rapidly growing megacities are another major problem, particularly in developing countries where massive slum areas and lack of social coherence render planning impossible and management difficult. UN estimates predict that almost the entire world population growth from the present six billion to eight billion in 2030, will be urban. This means an increase of urbanites from 2.8 billion to 4.9 billion, while the rural population is thought to remain stable at 3.2 billion. The most spectacular change will be in the now largely rural continents Africa and Asia where half the population will be urban in 2030. Of the 23 megacities (with more than 10 million inhabitants) expected around 2015, 19 are situated in developing countries. The megalopolis of the future risks to become a miseropolis if no appropriate action is taken! The problem can not be separated from the situation in rural areas: lack of opportunities there causes an exodus to the cities where often even worse problems have to be faced. Keeping rural areas economically viable and socially attractive is a major element in achieving a slower and better manageable growth of megacities.

SCIENCE AND TECHNOLOGY FOR DISASTER MITIGATION

The international scientific community, aware of the acute problems, has in recent years launched a range of large interdisciplinary research programs concerning the global environment, its changes and its significance for humanity. The relatively small, though rapid changes that have occurred in recent decades have already posed serious problems in many parts of the world and have made us aware of the great consequences of eventual lasting significant changes. The frequency and magnitude of extreme events and related disasters are increasing in the world's degrading and intensely used environments .

The present international disaster reduction research is focussed on a number of major problems, the responses to which are most urgently needed by society. Major topics include:

- The risks incurred in large urban/industrial areas where more than half the world's population is concentrated and where enormous property damage by natural and technological disasters can be expected. Hazard zoning, early warning and innovative engineering are important elements.
- The effects of changing land use and population patterns on disaster vulnerability. The increased population densities in critical zones and inappropriate land use systems are major factors.
- The responses required in densely populated coastal lowlands to face the expected sea level rise of the coming decades/centuries. Extensive, densely populated and food-producing areas are threatened.
- The study of Holocene and historical situations, trends and events in order to more precisely assess the effects of present environmental and land use changes.
- The disasters related to decadal climatic fluctuations, such as 'El Niño', the southern oscillation and the north-atlantic oscillation. Droughts, riverfloods and

mountain hazards are strongly affected by these and disasters, social stress, food shortage and economic losses will be the result.

- The global environmental dynamics on the basis of continuous monitoring by satellite data, observations on the ground, and modelling techniques. It has become clear that, in fact, our life support systems are at stake!

For each of these main research areas, all meeting specific needs of society, the scientific community has formulated research programmes and projects that are most urgently required. Studies of past trends and events should be coupled with mapping and monitoring of present situations and processes by modern aerospace technology and by data management through geographical information systems so as to assess the near-future developments. Matching the technologies with given disaster situations is not easy: there is a broad range of disaster situations to cope with and an equally broad range of aerospace observation methods (UN/IDNDR 1995, Verstappen 1995). For some hazards high spatial resolution or even 3-D information is required while others – affecting larger areas – can be dealt with by low spatial resolution data. High temporal resolution is needed for monitoring hazardous processes and situations. The greatest potential for these technologies is in disaster preparedness. Hazard zoning, guiding physical planning, and monitoring, leading to early warnings, are two major fields of application. A third type of application relates to emergency situations. The data then required are time-critical and thus all-weather capability of the sensor and timely satellite passes with short return intervals, are essential. The approaches to creeping disasters differ essentially from those used to cope with instantaneous disasters.

Creeping disasters comprise issues such as desertification and accelerated erosion with related increased flooding, existing non-sustainable land use systems and environmental changes. Global monitoring using low-resolution satellites is a major issue. The data so acquired may either be used directly to assess the present situation, or be entered in models for predicting future developments. The reliability of the models depends, of course, on the selection and proper weighing of the relevant factors on the one hand and on the accuracy of the data input on the other hand. Satellites providing systematic areal information of global distribution patterns give essential data input but should normally be complemented by traditional point observations for purposes of calibration.

The following brief review of current major global research programs illustrates the broad scope of the scientific efforts to contribute to the knowledge of life support systems and sustainable management of earth resources. The World Climate Research Programme (WCRP) comprises research on the global energy and water cycle (GEWEX), climate variability and predictability (CLIVAR), world ocean circulation (WOCE), the arctic climate system (SPARC), etc. The International Geosphere-Biosphere Programme (IGBP) comprises studies on biospheric aspects of the hydrological cycle (BAHC), global change and terrestrial ecosystems (GCTE), global atmospheric chemistry (IGAC), global ocean

flux (JGFS), land-ocean interactions in coastal areas (LOIC), past global changes (PAGES), global analysis and modelling (GAIM), data and information systems (DIS), analysis, research and training (START), etc. The International Human Dimensions Programme (IHDP) is concentrating at present on land use and cover changes (LUCC), but is developing several other activities.

The satellites used comprise low-resolution geo-synchronous satellites such as the US geostationary operational satellite (GEOS), the European METEOSAT satellite, the Japanese geostationary meteorological satellite (GMS), the Russian geostationary observation meteorological satellite (GOMS) and the Indian national satellite (INSAT) and sun-synchronous satellites in lower, near-polar orbits, such as those of the US NOAA-1 series, the Russian METEOR series and the Chinese FENG YUNG-1 series. Together they provide operational and high-frequency coverage of the entire earth at low cost. These data serve environmental monitoring and are within reach of most national and regional organizations concerned with disaster mitigation, particularly of the creeping category. Three, integrated, global observation systems are operational since 1992 and aim at bringing science and policymaking closer together. They do not only provide new data and information, but also make us better understand the current global environmental issues thus facilitating us to address them properly. They comprise:

- the global climate observation system GCOS,
- the global ocean observation system GOOS,
- the global terrestrial observation system GTOS.

Important new findings in this field are among others that climatic changes are not so much a fast response to radiative heating, but rather a slow response to heat exchange with the deeper oceans, which complicates models; and that global warming at the earth's surface is accompanied by global cooling of the stratosphere.

Rapid, instantaneous disasters evidently require entirely different research approaches and aerospace data. The International Strategy for Disaster Reduction (ISDR), the follow-up of the UN Decade for Natural Disaster Reduction (IDNDR), has launched projects related to seismic hazards in megacities, mountain hazards, floods etc. The research scenarios are geared to practical issues such as:

- investigation of the hazard susceptibility of the land,
- assessment of the vulnerability of society: endangered population, economic losses, etc.,
- quantification of the risk incurred in terms of life and property,
- preparation of hazard zoning and potential damage maps,
- monitoring potentially hazardous situations and processes,
- forecasting leading to early warnings,
- assessing emergencies and preparing disaster relief scenarios,
- pre-disaster and post-disaster planning.

The satellite data required is – with the exception of drought/famine and tropical cyclone disasters – high-resolution satellite imagery. LANDSAT (MSS or TM) and SPOT images are most commonly used, but also other space configurations such as the new American IKONOS, the Russian COSMOS and MIR, the Japanese MOS and the Indian IRS satellites may serve the purpose. The all-weather capability of radar satellite data from the European ERS, the Japanese JERS and the Canadian RADARSAT are very important, especially in case of river- and coastal floods, tropical cyclone disasters, etc. Radar interferometry, making it possible to measure vertical ground deformation with an accuracy of 1-2 cm, is a major technological advance with potential applications particularly for seismic and volcanic disasters.

The aerospace configurations and methods to be applied vary with the type of hazard. In the case of tropical cyclones emphasis is on monitoring the cyclone tracks using weather satellites and airborne/ground radar observations and giving early warnings. The system has been considerably improved in recent years and now functions satisfactorily. Hazard zoning on the basis of terrain configuration, frequency zoning on the basis of past events, disaster information systems, awareness raising and the construction of refuge platforms are elements of this success. Hazard zoning using remote sensing techniques is also a major technique in the case of mid-latitude coastal flooding, mainly by westerly storm surges. Emergency management, mapping the extent of flooded areas and post-event damage assessment require immediate radar surveys or infra-red photography. This also applies to river floods. Monitoring then is mostly by conventional means, using gauging stations upstream and automatic warnings when critical levels are reached. Hazard zoning using remote sensing and based on terrain configuration combined with frequency of occurrence, is a major theme. For optimizing warnings, however, precipitation data obtained from weather satellite radar recordings of rainfall intensity and duration and remotely acquired snow melt data of the headwater zones, are important. Early warnings hardly play a role in case of seismic disasters. Hazard zoning, based on site analysis (unstable slopes, wet areas susceptible to liquefaction, etc.) as a basis for physical planning and building codes, is a major issue. Radar interferometry from space, permitting recording of ground deformation with a vertical accuracy of 1-2 cm, is potentially a promising new tool for seismic disaster management in the future. Volcanic disaster reduction is rooted in the study of volcanic processes and landforms leading to hazard zoning maps for specific types of eruption activity. High-resolution aerospace images are an important tool.

CONCLUSION

The reality of environmental change and land degradation has, in the natural context, been generally accepted by the scientific community for some time. The concept of humanity as a driving force and as a recipient as well, is new, however.

Environmental hazard studies and disaster reduction efforts include physical as well as social elements and thus are in the sphere of interest of geographers. Their role in environmental research is, of course, not so much in systematic aspects such as biogeochemical flows and interactions, but above all in the regional differentiation and global aggregation of environmental niches and diverse human activities. For physical geographers there are three particularly promising fields of research in this area: analysis of present landforms patterns and earth surface processes, environmental geomorphology in the human context, including human impacts, and paleo-environmental studies related to Quaternary climatic changes as a means of predicting and assessing the effects of present and near-future trends of change. There is a range of technological research tools at their disposal to meet the present sustainability problems of the world.

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GEOMORPHOLOGY, ENGINEERING AND PLANNING

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ABSTRACT: The paper discusses ways in which geomorphology might be used by engineers and planners. They have always appreciated that a project should be based on a clear understanding of the geomorphology of the site of the proposed works. In many countries, following the lead given by Poland, it is now standard practice to take full account of the morphology and processes of the landscape during an investigation. However, many investigation teams tend to restrict the work to basic mapping and identification of features rather than a true understanding of the site in space and time. Not enough attention is paid to the residual effects of earlier landform change, the identification of inherited trends of change or the residual material conditions. Many projects do not have a long enough time scale to determine the frequency and magnitude of processes and the nature of risk even over the short design life of the project. Such deficiencies in practice are discussed and ways of improving the service are identified. Future work should be based on a full understanding of the conceptual basis for modern geomorphology. In the last thirty years there has been a revolution in the theoretical basis of the subject as well as a remarkable improvement in the technical capability. It is suggested that this should form a new conceptual basis for application to engineering and planning. Now that it is possible to discover fundamental information on how natural systems work it is negligent to carry out development or management of the surface of the Earth without basing the schemes on sound and available knowledge. It is the only home we have.

KEY WORDS: applied geomorphology, engineering geomorphology, event frequency and magnitude, geomorphological concepts, human well-being, inheritance, planning, risk, systems.

INTRODUCTION

In Poland and other countries of Central and Eastern Europe geomorphological mapping has been used for over 50 years as a planning tool and an essential guide to the development of engineering works. In Britain, however, it was not until 1972, on the recommendation of Professor P. G. Fookes, a leading British engineering geologist that K. Ainscow, partner of Rendel, Palmer and Tritton, Consulting Engineers, commissioned an experimental geomorphological map-

ping trial on the Taff Vale Trunk Road, South Wales. This became the first piece of detailed engineering geomorphological mapping in the UK to identify the location of trial pits and boreholes for a site investigation and take a project forward. In 1972, Ainscow, again guided by Fookes, then took the brave step of basing the preliminary alignment and reconnaissance investigations of the Dharan-Dhankuta road in Nepal on an engineering geomorphological assessment (Boyce 1995, Brunsten et al. 1978a,b, Brunsten et al. 1981, Cross 1982, Fookes et al. 1985). It identified the basic principles required for alignment selection in difficult mountainous country and a second survey in 1975 identified and mapped slope failures and drainage hazards and made qualitative recommendations for avoidance, stabilisation of slopes, earthworks and drainage provision for approximately 50 percent of the new road length. Many elements of the following site investigations, design and maintenance were based on the initial geomorphological model.

It is not generally known that the basis for the model was the stimulus provided by an innovative assessment of slope processes close to the project area in Darjeeling, by Starkel (1972). It is a pleasure to record the debt that British applied geomorphology owes to the Polish tradition.

These field experiments formed the starting point for my involvement in engineering geomorphology, a lifetime collaboration with many talented scientists and engineers, a blessing in the form of some of the best students available; together with R.U.Cooke, J.C.Doornkamp and D.K.C. Jones the creation of a professional company, Geomorphological Services Ltd, and involvement with truly exciting, innovative projects and Engineers (Brunsten et al. 1975a-c) such as the Bahrain Surface Materials Resources Survey (Doornkamp et al. 1980), the Review of Landsliding in Great Britain 1984-1987 (Department of the Environment), the Kotmale Reservoir Project in Sri Lanka (Doornkamp 1985) and the Channel Tunnel Portal (Griffiths et al. 1995, Brunsten 2002). In many of these cases the recommendation and initiative came from Professor Fookes. He is therefore recognised here as the godfather of British engineering geomorphology. It is a pleasure to couple my tribute to Leszek Starkel to a recognition of the contribution P.G. Fookes has made.

Despite the fact that these surveys were innovative and required courageous and influential action by the Engineer, a post-audit evaluation would reveal that two criticisms are made of the subject. First, engineering geomorphology is still not widely embraced by the engineering profession for the main reasons that it is seen as qualitative rather than quantitative, too judgemental, and not easily converted into engineering or even geotechnical design criteria. Secondly, the engineering content of the surveys could be increased if more attention were paid to the risk posed by the mapped geomorphology. In particular, the engineering would benefit from a greater consideration of the activity or frequency of process events and the sensitivity of mapped landforms and processes to construction effects, such as earthworks and drainage. There should also be impro-

ved interpretations of landform origin, definition of relict materials, assessments of slope stability and frequency, greater knowledge of sensitivity to change, greater confidence in prediction of ground conditions and evaluation of hazards.

The purpose of this paper is therefore to describe the nature of modern geomorphology and discuss how we might base our assessments on new approaches.

APPLIED GEOMORPHOLOGY

Geomorphology is the study of the forms of the surface of the earth, their origin, the processes involved in their development, the properties of the materials of which they are made and predictions about their future form, behaviour and status. Applied (e.g., Engineering) geomorphology is the application of this knowledge and the techniques involved, to the solution of a planning, conservation, resource evaluation, engineering or environmental problem (Brunsden et al. 1978).

The story of how ‘engineering geomorphology’ and ‘geomorphology for planners’ developed in the UK is well known and has been described many times (Brunsden 1985, 1996, Brunsden et al.. 1978, Cooke and Doornkamp 1974, 1990,

Table 1. Prospective clients of applied geomorphology.(Includes primary clients, who require and pay for development, and implementing clients, who design and construct the project and employ the specialist consultant).

Geologists	Require advice on specialist aspects of geomorphology, including mapping. It is not an easy task to record the shape of the landsurface nor to correctly identify landforms or relate form to process.
Developers	Require a cheap reliable service giving common sense advice and reassurance. They may also need the services of an expert witness through their solicitor in cases where they confront a land problem or wish to oppose a development.
Consulting engineers	Require independent advice on the viability of a scheme in relation to ground conditions, processes and impacts.
Planners	Require specialist advice on natural landforms, processes, materials, mechanisms, erosion, deposition.
Government	Require assessment of ground conditions and hazards which may provide a constraint on a planning proposal. A vital process is the avoidance of hazard by zonation and legislation rather than an expensive engineering solution.
Reinsurance	Require a scientific basis for policy advice on the management of the environment and the mitigation of hazard.
Lawyers	Require an assessment of natural disasters, ground and environmental conditions in order to assess vulnerability and risk and to respond to a request for cover, a claim for damage or loss. Require expert witnesses or other litigation guidance.
International agencies	Require a scientific basis for environmentally sensitive developments, evaluations of the viability of a scheme, expert panels to oversee or approve major developments or expenditure, rapid response teams in cases of disaster, technical workshops and educational programmes.

Doornkamp 1985, Fookes and Vaughan 1986, Goudie 2001, Griffiths 2001, Griffiths and Marsh 1986, Hutchinson 1983, 2001, Jones 1980, 1983). Earth management in the UK is an established part of the development and application of geomorphological theory. Geomorphological mapping and other techniques are now applied widely at all stages of a project including planning and feasibility, desk study, reconnaissance, site investigation, construction and post-construction phases, including operation and de-commissioning, and has the same full range of clients as engineering geology (Table 1).

Applied geomorphology can be divided into four kinds of work, problem identification, baseline surveys, innovative research and expert witness advice. These may require archive searches, walkover surveys, routine resource assessments, standard environmental statements, detailed mapping, monitoring, model building or new research depending on the stage of investigation. A personal view is that a geomorphological viewpoint is best applied at the outset of a project, but that the findings have relevance until completion because the aim is not just problem identification and the provision of the basic facts about a site but continuous monitoring to evaluate performance and especially to detect degradation of works or long-term responses such as pore pressure recovery.

EYES AND BOOTS – A QUALITATIVE SKILL

During the reconnaissance and site investigation stages there is a demand for that unique skill of being able to estimate past and present processes from the evidence of landforms and deposits. Because these are fragmentary this exercise is like reading a detective story that has the beginning, portions of the middle and the end missing. As all the great fieldworkers have pointed out there simply is no substitute for an ability to ‘read the ground’ or as Wooldridge (1956) put it ‘to feel the ground through the sole of your boots’. I share here, with Hutchinson (2001) the view that although the first insights are often judgemental they have great worth. It is my view that the best way to achieve this is to work as part of a ‘Geo-Team’ to establish a first stage geomorphological input to the main geological model, using the principles described by Fookes (1997). We will then be in a strong position to go on to detailed and quantitative earth science including measurement, monitoring and sophisticated modelling.

It is emphasised that an historical approach is necessary if we are to understand the time and space controls on a project. The challenge is complex because we need to evaluate the probabilities and trends of landscape change. A planner, engineer, manager or earth scientist therefore must be widely educated, flexible, experienced and free thinking. Because the environments, materials and processes are so varied and can be anywhere on Earth and yet we must maintain the same high standards, it is essential that we professionally apply clear generally applicable methodologies, codes of practice, tried and tested standard methods. That is why the Geo-Team approach and ‘total geology’ (Fookes et al.

2000) are so important, in which we can share knowledge and nurture the inexperienced. That is why we must also have supporting (chartered) institutions and a growing volume of case histories of both best and failed practice.

THE CHANGING GEOMORPHOLOGICAL FRAMEWORK FOR APPLIED STUDIES – 1970-2000

British geomorphologists in the early 1970s were trained as geographers because geomorphology in the UK was primarily carried out in those departments. The nature of geography, which is to synthesise data from many sources made them ideal scientists to contribute to the feasibility, walkover and site investigation phases of projects. The geographical skills which became valuable were to be able to appreciate the significance of spatial and temporal patterns and events; to appreciate how variables change in importance with the scale of the problem and an ability to use maps and geographical information systems and information data banks to map and analyse the development of the landforms. It is therefore not surprising that the basic skills used at that time were morphological mapping, surveying, the interpretation of air photographs and satellite images, soil and vegetation surveys. The dual training with human geography also provided an appreciation of the politics and agendas of the financial and policy formulation world that underpins all practical schemes. This meant that the needs of the client and the imperative to translate the technical aspects of the subject into terminology suitable for the layperson became a characteristic of geomorphological reports.

Fortunately, during the 1960-70 this conceptual framework was changing from a long term, deductive view of landform change as an extension of historical geology, with the purpose of constructing landform change and denudation chronology models, to a system based study of process mechanisms and rates. This approach emphasised a balance between work and load, force and resistance and required precise mapping of the landscape, monitoring, data collection, identification of error and the statistical techniques of systems analysis. Technical proficiency and a need to measure processes and materials became common practice and it was this that brought geomorphologists into contact with the deterministic modelling of engineering and planning.

During the 1970s the need for technical competence introduced the use of many advanced techniques so that today there is a capability to remotely observe the earth using a variety of space and air-borne platforms; to collect data using remote data loggers; to date events using sophisticated scientific methods; to analyse properties and to simulate and model using fast and powerful desk-top computing facilities. There has been an improvement in inventory, mapping, calibration, evaluation of hazard, susceptibility, vulnerability and risk calculations and stability studies. There was also an awareness of the importance of engineering and geological performance standards, codes of practice, legislation, envi-

ronmental policy acts, and insurance restrictions. These developments meant that geomorphology is now able to offer a rigorous, quantitative service to a client.

The explanation that follows is a personal view of a modern framework for the subject. It is, however, only a summary of the parts relevant to practice; for a full explanation see, Brunsden (1990,1993a,b, 1996, 2001) and Brunsden and Thornes (1979) which contain full references to the development of modern geomorphology.

It is suggested that the following propositions might be used as a conceptual basis for the required predictions and calculations. They are based on the view that the task of the geomorphologist is:

- to provide the Geo-Team with the specifications of the process-response systems in the project area,
- to describe the preparation processes, inherited trends and triggers that govern change,
- to identify the processes that are currently forcing change,
- to isolate the sensitive areas of the landscape,
- to predict the nature of system behaviour in the future.

A WAY FORWARD

The complexity of many projects and the high standards of the engineering-geosciences profession require that a holistic approach is adopted to site investigation. This paper suggests that, where possible, an integrated Geo-Team be employed. To contribute to this the approach to modern geomorphology outlined above can be summarised for practical purposes into eight tasks.

- The structure of the system: describe the landforms, structure, materials and climate.
- The process-response regime: identify and monitor the processes; analyze the records, map and monitor the behaviour.
- The event regime: establish the nature of the events that occur over time and space that might affect the project, including their frequency, magnitude, duration and sequence. Describe the episodic nature of landscape change, the effective events and the formative events.
- Preparation: identify and assess the effect of both natural and human time dependent processes that prepare the system for change and which occur during the lifetime of the project.
- Complex cause-the forcing functions and changing controls: assess the likelihood that one of the controls of the system might change during the lifetime of the project and describe the changes that must be catered for in design. This includes changing climate, rising sea level and storminess. Map the progress of changes already in progress.
- Inheritance and sensitivity to change: determine the trends of change and

waves of aggression already attacking the system, discover and define the sensitive areas and determine the internal and external thresholds at which failure, reactivation or deformation will take place.

- Behaviour: record the behaviour of the system, beach erosion and deposition, foreshore lowering, cliff recession, vectors of slope movement.
- Risk: estimate the risks to the project.

THE STRUCTURE OF THE SYSTEM

The general approach adopted is to recognise that the Earth's surface is composed of process-response systems that are specified by the stress fabrics given by the tectonic history and setting, the properties of the rocks, superficial deposits, landform geometries, climates and biomes arranged in patterns across the earth's surface in an hierarchy of landscape types and scales. The great value of this approach from the applied point of view is that the stress fabrics, rock types and morphometric properties are measurable and can be mapped using desk study, morphological or engineering geological mapping methods. They can be conveniently arranged in suitable classification schemes such as landform types, land systems, regions, drainage basins or geological models. The most useful products are then early warning maps, preliminary hazard assessments, identification of problem areas for future detailed mapping and the creation of the first phase ground models by the 'Geo-Team.'

THE PROCESS EVENT REGIME

The initial specifications of each tectono-climatic area also determine the processes, the materials, the forces and resistances, flux and the rates of erosion and deposition of the landscape. Operating within each area there is a process regime (weathering, mass movement, fluvial, glacial, periglacial, aeolian, coastal). Each regime is characterised by an hierarchy of events distributed as a sequence in time and space and described by the frequency and magnitude, duration and sequence of process events.

It is important to recognise that the process scales can change over time. The change of energy, mass or process rate may be sustained at a new level if a small but permanent shift in the controlling variables takes place (e.g., land clearance or mining) or return near to their initial state if the energy pulse does not exceed the elasticity of the system. Normally these events are tectonic or climatic in origin, spatially and temporally restricted.

River and coastal engineers, of course, understand recurrence interval graphs, stage and discharge records and wave height diagrams that occur on a 'design life' time scale but are less familiar with understanding the effect of events that occur or have occurred on a non-engineering geological frame-

work. The challenge for the Geo-Team is to identify and interpret the engineering significance of past events for all processes and at all the time scales that may occur during the lifetime of the scheme.

THE EVENT REGIME

Each process has internal threshold values or external trigger values at which the process becomes effective and sediment transport and landform changes are mobilized. Perhaps the easiest example to illustrate this is the intrinsic strength thresholds of hillslopes. Some are the original properties of the system specifications (e.g., the peak strength mobilized during first time failure); some are 'prepared' properties of the system (e.g., the fully softened strength); some are the 'failed' properties after a slope movement (e.g., the residual strength on relict shears that form the basis for reactivation). There are numerous unfortunate cases where failure to recognize the different threshold values has led to serious failure. It is important to recognize that they can themselves evolve as a result of time-dependent evolution of the system resistances. The events that cross these internal thresholds are sometimes called the effective or triggering events, such as the rainfall that triggered a landslide but this is a simplistic view of landscape change. It is essential that direct causes are understood in the context of long-term preparation and the inheritance trends of the system. These can be evaluated from process-response observations or calculated from ground models. The search for such values is the essence of research on natural hazards.

FORMATIVE EVENTS

In addition to the slow operation of the most frequent processes, each landform assemblage is also subject to those events that control the evolution of the form of the land. The formative event may or may not carry out the most work over time but it is the event that is responsible for creating a form that persists for long periods, despite the modifying actions of more frequent events.

The frequency may be such that the form created, often in a short time, persists until another formative event occurs. It often has a lifetime longer than the 'creation' time (Brunsten 2000). This fact determines that different forms and deposits may persist for a long time and that there will be a strong juxtaposition of forms of different age and behaviour in the landscape.

EVENT SEQUENCES

The series of events that affects each landform assemblage is unique in the detail of occurrence and sequence. No two systems receive exactly the same number, sequence, frequency, duration and magnitude of events although in 'uniformitarian' terms they are subject to the same processes and the results

may 'average out' over time. It follows that there will be a spatial variation in the response forms of neighbouring systems, solely because of their history. A river valley may not record exactly the same changes or landforms as its neighbour and this effect may magnify or diminish with time.

The exact sequence will be very important in terms of whether an effective event precedes a weaker occurrence (all the available work will have been done and there will be a need for a recovery period during which the preparatory processes provide further sediment for transport. Crozier et al. (1990) describe such landforms as suffering from 'event resistance', Brunsten and Thornes (1979) called them over-relaxed. A very good example is a debris flow system that has suffered a big storm, evacuated all the material and requires a period of weathering before it has enough debris to flow again.

COMPLEX RESPONSE

In addition to the event responses noted above, as a system responds to an event, the forms will relax to a state that is in harmony with the forces being applied. Then 'for any set of environmental conditions, through the operation of a constant set of processes, there will be a tendency over time to produce a set of characteristic landforms' (Brunsten and Thornes 1979). This involves internal feedback mechanisms and there are four important responses. If the event does not exceed the tolerance of the system no significant change will take place. If the system is self-regulatory it will adjust in such a way that it will tend to minimize or 'damp-out' the effects of the disturbance. If the disturbance exceeds the ability of the system to return to its previous form then the system will hunt for a new characteristic state and form and to a process rate that can be maintained at a new level of geomorphological activity. There will then be a new unity of landscape and a new interdependence of process and form at all scales. This idea, that there is an underlying order in landscape complexity, is the basis of landform mapping and classification and of its use in applied geomorphology. If the impulse of change reinforces an existing trend then the magnitude, type and rate of change will also alter. Major change will occur if the induced change crosses a critical structural specification of the system. If the event is so big that it completely erases the system then a 'geo-catastrophe' may be said to have occurred and the Earth's surface is re-specified. Such changes may be very complex at all temporal and spatial scales (Schumm 1973, 1977).

DEFINITIONS OF TIME

It follows that we can now redefine change in time. The length of time that elapses between impulses of change is known as the recurrence interval. The time required for the system to 'notice' that an impulse of change has been applied is the reaction time. The time required for the system to respond and to reach a new characteristic state is the relaxation time. The characteristic state time is the

time during which a state and form persists as a diagnostic element of the landform assemblage. This is also called the landforms 'lifetime' (McSaveney and Griffiths 1988). The transient form time is the time during which the landscape attempts to reach a slowly changing form but is continually being interrupted by another impulse. The landforms are then rapidly fluctuating and never attain a new, stable, slowly changing state. Such forms are called transient landforms.

One of the most important aspects of this view of time is to draw attention to how long it takes for a system to recover following a disturbance such as land use change, beach supply depletion, groundwater extraction or channel change a after flood. Available evidence suggests recovery times of 10^3 or 10^4 years for hill slopes, 10^2 years for beaches and 10^1 years for channels. Human beings should, perhaps, be aware of these timescales before disturbing an environment!

Change through time therefore becomes a series of responses to all of the events experienced by the system. In the field this directs attention to the fragmentary landforms and material relicts produced by past processes that indicate the possible magnitude of what might happen again in the future. It also suggests that a search be made for events that occurred some time in the past but whose effects are still affecting the system behaviour. Herein lies a deeper understanding of natural hazard and 'risk'.

PREPARATION

Each regime also sets in motion time-dependent changes that progressively alter the balances between forces and resistances by altering the material properties, the geometry of the system, the water regimes and the time to occurrence of landscape change (Tables 2-3). Preparation for change occurs throughout the life of a landscape and is expressed through the evolving internal thresholds referred to above. They include the chemical and physical changes of weathering; the progressive softening of materials or the growth of armours and cements; changes in permeability and groundwater levels; consolidation and stress history; loading and unloading and slow geometrical changes to relief and gradient due to erosion or even changes to the urban fabric.

Both the processes and the changes can be measured and monitored over time although some changes are very slow. It is important to distinguish whether the system is becoming more stable and absorbing the effect of the changes by some negative feedback mechanism or whether the change is a reinforcement mechanism leading to rapid and irreversible change.

COMPLEX CAUSE-FORCING FUNCTIONS AND CHANGING CONTROLS

The concept of 'complex response' also generates the fundamental proposition of 'complex cause' that, 'landform change takes place as states of equilibrium, stability or tranquillity are upset by complex episodic changes to the envi-

Table 2. Natural preparatory and triggering processes in geomorphology (Brunsden 2001).

System specifications	Morphological control	Geological control		Hydrological control	
	Erosion-transport processes	Structure	Lithology	Ground water	Surface water
Preparatory process	Undercutting Headcutting Sediment transport Deposition	Stress relief Joint Formation Softening	Weathering regolith and colluvium depth	Water table levels	Infiltration Runoff freezethaw
Forcing functions	Environmental/climatic change Slope ripening Anthropogenic changes				
Specification change	Geometric Change Angle Height length	Permeability strength	Regolith depth chemical and physical properties	Flow net pore pressure values	Infiltration capacity runoff coefficient ice growth
e.g., Direct trigger	Critical slope	Critical strength	Critical depth	Critical pressure	Critical volume
e.g., Mass movement response	Debris slide	Rock slide	Debris fall	Rotational slide	Debris flow

ronmental controls’ (Brunsden and Thornes 1979). In addition to the disturbances applied to a system by the normal events, each dynamic system is subject to changes in the controlling causal parameters. The main point being that all sites, forms and processes will be subject to the episodicity and rhythms of Earth that will alter the event sequences and magnitudes (Brunsden 1990).

Examples of forcing functions that affect the current geomorphological systems are, tectonic movements, active faulting and the seismicity, the varying sea levels experienced during the lifetime of the landscape, the perturbations that arise from changes in the environment (climate-vegetation), and the occurrence of rare, formative, catastrophic events that may change the directions and magnitudes of evolution. For engineering and planning it is very helpful to determine whether a site has been or is being affected by low sea levels, by sea level rise or changing climatic frequencies that cause increased erosion activity.

It is clear that the forcing functions are a direct cause of the diversity and complexity of the earth’s surface. These impulses are episodic and complex in nature at all scales. Therefore the changes to the systems will be episodic and complex (Schumm 1973, 1977). If, following Fookes (1997, Fookes et al. 2000) we intend to build ground models some attempt should be made to build in the scales of these episodically acting controls.

Table 3. Anthropogenic preparatory and triggering processes in geomorphology (Brunsten 2001).

System specifications	Morphological control		Geological control		Hydrological control	
	Construction		Structure	Fabric	Ground water	Surface water
Preparation process	Cut/fill	Blasting vibration	Pile Rock anchor Pin Retain Reinforce	Chunam Shotcrete Compact Grout	Tunnel Drain Supply Sewer	Catch Store Drain Supply
Forcing functions	Planning Dynamic growth of city and infra-structure Maintenance and control Broken services					
Specification change	Load Unload Slope form	Permeability Strength	Bulk strength voids	Erosion Strength voids	Height Pressure Internal erosion Weep hole Efficiency Chemistry	Volume Erosion
e.g., Direct trigger	Critical stress	Critical strength	Critical stress and strength			
e.g., Mass movement response	All forms of slide	Rock falls and slide	All forms		Rotational Non-circular	Debris flow

INHERITANCE AND SENSITIVITY

The engineering site and situation is created by the operation of the system over time and the dynamic state is determined by the current activity and future change. It is a system with a history as well as a current state. Many systems have been running a long time, have suffered stresses, have failed, been repaired and are considerably worn (Di Nocera et al. 1995). Many of the most serious problems faced by the 'engineer' arise from this fact and many of our greatest mistakes have been result of a failure to recognise the signs of age and wear. In our assessments of the current properties of the site it is essential to recognise the inheritance.

It is helpful to place emphasis on the trends of change that are actually in progress, e.g., tendency to erode, aggrade or move the location of activity. In 1971, Bjerrum described the morphological discontinuity or erosion front, where quick clay landslides were eating into a marine platform, following stream incision in response to the tectonic uplift as a wave of aggression. The point reached by the aggression should be regarded as the most fundamental of all

morphological points to be brought to the attention of the engineer. For example, Bjerrum (1971) provided a solution to the control of the occurrence of quick clay slides by armouring the streambed and preventing the progress of the energy pulse. The stream banks were not eroded nor were the streambeds deepened and so the quick clay could not flow away from its store. The incision of a river into an oldland or the retreat of a sea cliff into a plateau are other examples.

It is also useful to know the likelihood that a system will 'fail' or change. This is known as its sensitivity and involves a knowledge of the inherited strengths and weaknesses, barriers to change, location of change and stability (Brunsdon and Thornes 1979, Thomas and Allison 1993, Brunsdon 2000, Thomas and Simpson 2001).

A useful measure of sensitivity is the factor of safety as measured by the ratio between the disturbing and resisting forces. The disturbing forces are provided by the application of energy from the specified tectonic, climatic, biotic and anthropogenic controls on the geological, hydrological and morphological framework of the system. The resistances are complex and can also be thought of as barriers to change. These are:

- Strength resistance. The properties and structure of the materials of the system.
- Morphological resistance. The distribution of potential energy as elevation, made available as a function of slope angle and relief i.e. the energy of position. Sensitivity is the ability of the system to mobilize this energy.
- Structural resistance. The design of the system, components, controls, thresholds, and the way they are linked.
- Locational resistance. The distance to a propagating source of change, e.g., Centre of uplift; distance to base level; closeness to an undercutting stream.
- Transmission resistance. The ability of the system to transmit change. A high density of flow lines (e.g., rivers, joints) and good linkage helps efficiency. Elements that are coupled together allow easy change e.g., the linkage that allows the undercutting of a slope by a stream, to translate sea level lowering along a river and eventually to cause a landslide on the watershed.
- Filter resistance. The way in which a system absorbs or removes energy from a system. Energy diffusion across area and the use of energy absorbers such as beaches and energy barriers such as waterfalls are the main controls.
- Inherited resistance. The resistance or state of a system also varies because of the inheritance including the degree to which it has recovered from an application of stress.

BEHAVIOUR

The specifications of the original system (structure, process domain, event regime), the changes made by the time dependent preparation processes and the inheritance of variably sensitive areas, trends of change in operation, the current

forcing functions of environmental change, the process and residual thresholds determine the behaviour of the current systems at a site. Normally, at the start of a job, a partially used, weathered and worn residual landscape is inherited. As time passes the systems continue to operate and it is this behaviour that must be managed. The behaviour may be stable, transient between different regimes or reactivating. There will be varying degrees of activity, rates, deformation styles, sediment fluxes and reactivation of dormant episodic systems.

One of the most innovative achievements of applied geomorphology for planners has been the development of Ground Behaviour maps (Moore et al. 1991, Moore et al. 2000). The idea is capable of considerable development and can be extended to other processes.

RISK

Risk is usually specified using standard mapping of sensitive sites and analysis of frequency-magnitude statistics. Recent work also uses expert judgement, Delphi panels and subjective probability assessments derived from event trees (Roberds 1990, Hall et al. 2000). These risk assessment methods offer the potential to quantify the effects of the uncertainty inherent in the operation of the processes. They aid and improve decision-making by allowing consideration of a range of possible scenarios and consequences, each with different probabilities of occurrence. It is an iterative process in which the issues which contribute significantly to the total risk are identified, and the less important issues are screened out in a systematic and rational manner. Risk assessment asks questions about what could happen? Cause, chance, loss, damage, management of problems, and reduction of risk are key considerations.

Geomorphology can contribute to risk assessment by assessment of the probability of a particular type of event or level of movement occurring and by providing an indication of the potential impact of a particular type of event on the elements of risk.

NEW INSIGHTS

The approach outlined above yields several important insights to the 'old game' of 'eye for country.'

- The use of detailed geological and geomorphological mapping to describe the structure of the system is a successful guide to the identification of the individual system components, landslide subsystems and the interrelations between them and enables a very efficient site investigation to be designed.
- The wide spatial area studied by the Geo-Team draws attention to the natural patterns and provides a sound understanding of the situation of the project area within the controlling process-response systems.
- Preliminary ground models can be constructed from the mapping programs and used as a basis for detailed ground investigation. The results can then

be extrapolated with considerable confidence and the solutions are more cost effective because there is more understanding of system behaviour.

- The knowledge of the episodocity of the events, their effectiveness, changing frequency and formative threshold values give great confidence to the open forum panel discussions for risk analysis because the probabilities could be calculated with some scientific basis.
- A great improvement in understanding derives from the wide range of time and space scales considered by the team. This ranges from long term structural geology and lithological considerations, to Quaternary events, Holocene environment and sea level changes, millennial scale evolution models, centennial scale archive and cartographic reconstructions, decadal knowledge of climatic events and process frequency and daily or hourly water and deformation monitoring. Space scales range from the tectonic setting to the contents of one borehole or the aligned particles of one shear surface. This understanding gives great economies of scale because each investigation throws light on other investigations at different scales and makes the integration of the sub-disciplines complete. In particular the depth of understanding of space and time ensures that the site is understood in a wide evolutionary context. This gives an appreciation of the effect of proposals on neighbouring systems so that extrapolation of results can be made with confidence.
- The emphasis on the inheritance of the system promoted understanding of the sensitive forms created in the past (e.g., relict shear surfaces and residual strength conditions); the trends of evolution and the action of the forcing functions that alter the aggression of the system.
- The recognition of successive waves of aggression and of the inter linkages between the system elements enables the construction of event trees based upon detailed knowledge of the sediment and water cascades. The work of the open forum panel can therefore be based on a clear scientific rationale and the risk analysis can be used with confidence. The combination of these factors together with knowing the exact system boundaries of each behaviour unit then forms the basis for cost benefit studies based on the area first affected and expansion during time. The cost-benefit calculations therefore have a theoretical underpinning that stems from Geo-System behaviour considerations rather than standard extrapolations from past performance shown on historical maps. There is thus a direct application of the work of the Geo-Team to the costing of the project.

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As all manuscripts will be reviewed, therefore, authors are kindly requested to provide two hard copies of the complete text (main text, abstract, references, tables, illustrations (one set of illustrations may be in xeroxed form) and one 3.5" diskette in MS Word.

The manuscripts should be arranged in the following order. First sheet: title, full name of author(s), affiliation, full postal address, e-mail address (if possessed). Second sheet: abstract of no more (no less) than 100 words, key words (3-10) at the end of the abstract. Subsequent sheets: the main text of about 20-25 pages double-spaced (in A4 format). Then on separate sheets: acknowledgement (if desired), notes, references, tables, illustrations, captions to illustrations.

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References should be indicated in the text by listing, in parenthesis, the author's name followed by the data of publication, eg. Kowalski (1995) or Kowalski at al. (1996) if there are more than two authors (Note: all the authors should be listed in the references)

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