Water and sediment discharge from glacier basins: an arctic and alpine comparison

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Abstract This paper compares the discharge and the dissolved and suspended sediment transport from two glacier basins: the Haut Glacier d'Arolla basin, Valais, Switzerland and Austre Brøggerbreen basin, Svalbard. Information from the two basins has been split into subperiods to represent different components of the ablation season. Three differences are identified between the magnitude and timing of flow and sediment transport components: (a) in the alpine basin, discharge follows a distinct seasonal sequence which is driven by the balance between direct solar radiation input and reflected radiation from a glacier surface of progressively changing albedo. Although a similar ablation season sequence is identifiable for the arctic basin, it is relatively subdued and heavily modified by the influence of advected energy on melt rates; (b) the solute concentrations in the meltwater of the rivers in both environments (as described by continuous monitoring of electrical conductivity), also shows a distinct pattern through the ablation season, but the diurnal variations observed in the alpine environment, which have been attributed to either the mixing of runoff components or variability in the nature of the weathering environment, are not present in the arctic record; (c) the widely-reported "exhaustion" in suspended sediment supply to alpine proglacial streams over the summer ablation season is not evident in the arctic basin. Indeed, increased supply of suspended sediment may occur through the arctic ablation season.

INTRODUCTION

Whilst much research has been focused upon the hydrology of alpine glaciers, arctic glacier hydrology has received considerably less attention. The arctic environment provides great potential for hydrological study because there is a complex range of potential ground and ice temperature conditions (e.g. Blatter & Hutter, 1991) which have additional and complex influences on hydrological processes in addition to the seasonal movement of the snow line, the associated change in glacier albedo, and the resultant change in the amount and routing of meltwater to the proglacial river that supports the typical seasonal hydrological pattern in alpine glacier basins (Röthlisberger & Lang, 1987).

This paper contrasts time series of discharge, suspended sediment concentration and electrical conductivity from the proglacial rivers draining a warm-based alpine glacier and a polythermal sub-polar glacier of similar size. Information from two ablation seasons for each basin may well be unrepresentative of the longer-term character of the basins, but the differences in the records are sufficiently marked to illustrate some of the impacts of climatic and ice temperature gradients on the regime of proglacial rivers.

THE STUDY AREA AND DATA COLLECTION

Austre Brøggerbreen (an arctic glacier) is located some 3 km to the southwest of Ny-Alesund, northwest Svalbard. It has an area of approximately 11.7 km^2 rising from approximately 50 m at the snout to 720 m above sea level at the watershed. Meltwater drains to Kongsfiord via the Bayelva River. The bedrock consists of Devonian, Carboniferous and Tertiary sedimentary rocks which include limestone, dolomite, shale and red sandstones. The glacier has retreated considerably over the last 100 years. Maximum ice thickness is currently approximately 130 m and the glacier base is polythermal, being warm beneath ice thicknesses in excess of 70-80 m and cold beneath shallower ice (Hagen *et al.*, 1991). The catchment receives continuous daylight from 18 April to 24 August, although frequent periods of coastal fog and high cloud cover strongly influence the receipt of direct solar radiation (Hodson *et al.*, in preparation).

Haut Glacier d'Arolla (an alpine glacier) is located in the Val d'Herens, Switzerland, and has an area of approximately 6.3 km^2. The glacier ranges in altitude from 2560 m at the snout to 3838 m at the watershed. The glacier has been retreating in recent years, is warm-based and has a maximum ice thickness of approximately 180 m (Sharp *et al.*, 1993). The catchment is underlain by schists and gneisses of the Arolla series. Snow and ice melt is strongly influenced by diurnal variations in incident solar radiation (Gurnell *et al.*, 1992a).

This paper reports on variations in the discharge, electrical conductivity and suspended sediment concentration of the proglacial rivers draining the two study catchments over two ablation seasons: 1989 and 1990 for the Haut Glacier d'Arolla catchment and 1991 and 1992 for the Austre Brøggerbreen catchment. The field monitoring periods for the Haut Glacier d'Arolla extended from early June to early September in both years, whereas the shorter field seasons for the Austre Brøggerbreen extended from late June to mid-August in 1991 and to early August in 1992. The timing of the monitoring periods is summarized in Table 1. They include the phase of snowmelt from the glacier foreland and extend well into the phase dominated by melt from the glaciated portion of the catchment in all of the four field seasons. In both catchments the electrical conductivity and suspended sediment concentration of the proglacial river were monitored within 200 m of the glacier snout. Electrical conductivity and water temperature were monitored using a pHOX 57 (Mk 2) conductivity and water temperature recorder. The conductivity data were recorded at the ambient water temperature which was always in the range 0-2°C. Suspended sediment concentration was estimated by filtration of water samples collected at two and three hourly intervals in Switzerland and Svalbard, respectively, using an ISCO automatic pump sampler. The

Table 1 The timing of the subperiods identified for each of the four ablation season data sets.

	Period	Start date	End date
Haut Glacier d'Arolla 1989	1	1 June	8 June
	2	9 June	15 June
	3	16 June	8 July
	4	9 July	6 August
	5	7 August	31 August
Haut Glacier d'Arolla 1990	1	29 May	19 June
	2	19 June	2 July
	3	2 July	27 July
	4	27 July	13 August
	5	13 August	26 August
Austre Brøggerbreen 1991	1	27 June	3 July
	2	4 July	14 July
	3	15 July	5 August
	4	6 August	14 August
Austre Brøggerbreen 1992	1	24 June	3 July
	2	4 July	10 July
	3	11 July	16 July
	4	17 July	22 July
	5	23 July	28 July
	6	29 July	3 August

water samples were filtered through Whatman 40 (Switzerland) and Whatman 0.45 μm cellulose nitrate (Svalbard) filter papers. In Switzerland, these suspended sediment concentration determinations were used to calibrate a continuous turbidity record, monitored using a Partech 7000 series model 3RP suspended solids monitor fitted with a single head probe (Gurnell *et al.*, 1992b), to derive an hourly time series of suspended sediment concentration estimates. Lower suspended sediment concentrations and consequent interference from ambient light precluded a similar approach in Svalbard.

In Svalbard, discharge was monitored at the same site as electrical conductivity and suspended sediment concentration by developing a discharge rating relationship between current-metered discharge estimates and water stage monitored by pressure transducer. Although a relatively stable river section was used for discharge monitoring, it is likely that errors associated with the discharge estimates are at least 10% (Hodson, 1994). In Switzerland, hourly determinations of discharge were provided by the Grande Dixence hydropower company who monitor flow at a rectangular sharp-crested weir structure located approximately 1 km from the glacier snout. There are no significant inputs of water to the river between the water quality monitoring site and the Grande Dixence gauging station.

Thus, in summary, hourly determinations of electrical conductivity and discharge were available for two field seasons from both glacier basins. In addition, hourly estimates of suspended sediment concentration were available for two seasons for the Haut Glacier d'Arolla, and three hourly determinations were available for only one season (1992) for the Austre Brøggerbreen basin.

DATA ANALYSIS

The proglacial discharge, electrical conductivity and suspended sediment concentration time series monitored during the four periods were subjected to similar analyses in order to allow comparison of the two study catchments. Each field season was split into "hydrologically-meaningful" subperiods which could be analysed and compared. In the case of the Haut Glacier d'Arolla, this subdivision was achieved by identifying periods of similar magnitude and diurnal variability in discharge (Gurnell *et al.*, 1991; Gurnell *et al.*, 1992a), which is a measure of the amount, storage and routing of meltwater (Gurnell, 1993; Röthlisberger & Lang, 1987). These latter factors affect the degree to which the meltwater has access to sources of suspended sediment and solutes. Five subperiods were identified in each of the 1989 and 1990 data sets (Table 1). In the case of the Austre Brøggerbreen catchment, subperiods were isolated on the basis of patterns identified within time series of hourly observations of solar radiation receipt, wind direction and air temperature (Hodson *et al*, in preparation). This approach was adopted because of the substantial influence of advected energy (represented by the wind direction and air temperature) on melt rates, in addition to the influence of direct solar radiation receipt. Four and six subperiods were identified for the 1991 and 1992 field seasons, respectively (Table 1). Splitting ablation season river flow data into subperiods prior to analysis is consistent with the approach adopted in analysing similar alpine glacier-basin data (Jensen & Lang, 1973; Lang & Dayer, 1985). The following three groups of analyses were applied to the data sets for each subperiod: *firstly,* the lags or leads between the time series were estimated; *secondly,* suspended sediment concentration and conductivity-discharge relationships were estimated between the series using linear regression analysis after lagging the series to their best match position; and *finally,* Box & Jenkins (1976) ARIMA models were estimated for each of the time series.

Table 2 provides a first indication of the changing relationships between the monitored time series at a seasonal and sub-seasonal level and also some differences between the relationships in the two study basins. In all cases the discharge and suspended sediment concentration series were $\log_{10}$ transformed to stabilize the variance in the individual series and to generate the most linear relationship between the series. The best match position between each pair of time series was then identified from the cross correlation function estimated between the series.

Table 3 lists the parameters of the simple linear regression relationships estimated between the lagged electrical conductivity or suspended sediment concentration series and the discharge series. A subdivision of periods 3 and 2 in 1989 and 1990, respectively, at Haut Glacier d'Arolla was necessary prior to estimating the relationship between electrical conductivity and discharge.

Table 4 lists the form and parameters of the ARIMA models estimated for each of the time series. The approach adopted for model identification, estimation and diagnostic checking is described in Gurnell *et al.* (1992a). In each case, the parameters of the models are all significantly different from zero ($P < 0.05$) and the residuals represent white noise once the time series has been filtered by the ARIMA model (i.e. there was no identifiable temporal pattern remaining in the residuals as indicated by the autocorrelation function (acf) and partial autocorrelation function (pacf) for the residual series and by a Portmanteau χ^2 test applied to the first 24 autocorrelations of the residual series (acf).

RESULTS

Two major trends can be identified from Table 2: the lags/leads of the alpine series are much shorter than those of the arctic series; and there is a clear seasonal trend in the lags/leads for both environments. The difference in lags/leads suggests major contrasts in the functioning of glacier hydrological processes in both environments. In the alpine system, it has been suggested that meltwater is routed through a series of reservoirs whose size and residence time change throughout the ablation season (Gurnell, 1993). The residence time of meltwater in contact with rock debris influences the solute load of the meltwater, and an inverse relationship between solute concentration and discharge is induced by the mixing of meltwaters draining reservoirs in different locations and with different residence times (e.g. Brown & Tranter, 1990). The best match positions indicated by the cross correlation functions between the series (Table 2) all indicate a negative relationship between electrical conductivity (a surrogate for varying solute concentration) and discharge, but the longer and more variable lags from the arctic basin suggests that other major influences may be operating. Whilst there are very small lags

Table 2 Best match position between raw time series (identified from cross correlation functions).

Input	Output	Time period: 1	2	3	4	5	6	Direction of correlation
		(Time (h) of best match)						
Haut Glacier d'Arolla: 1989								
$\log_{10} Q$	cond	*	−1	−1	−1	−2		−
$\log_{10} Q$	$\log_{10} S$	*	0	0	+2	0		+
Haut Glacier d'Arolla: 1990								
$\log_{10} Q$	cond	0	0	−1	−1	−1		−
$\log_{10} Q$	$\log_{10} S$	*	0	0	0	0		+
Austre Brøggerbreen: 1991								
$\log_{10} Q$	cond	*1	+2	+3	+4			−
Austre Brøggerbreen: 1992								
$\log_{10} Q$	cond	*	+2	0	+3	+2	0	−
$\log_{10} Q$	$\log_{10} S$	*	−6	−5	−5	−2	−1	+

Q – discharge; S – suspended sediment concentration; cond – electrical conductivity.
* No clearly defined best match position.
*1 The time series for period 1, 1991, are short with many missing values.

between the positively correlated suspended sediment and discharge series for the alpine basin, the suspended sediment series leads the discharge series in the arctic basin, and the lead decreases through the ablation season. This could suggest a difference between the source areas of suspended sediment and discharge in both systems, with the suspended sediment sources being located closer to the monitoring site, (but with some convergence in the locations of the sediment and meltwater sources as the ablation season progresses) in the arctic system.

The linear regression relationships in Table 3 are estimated after lagging to the best match positions. Lagging removes some of the hysteresis in the relationship between the series, but significant hysteresis remains so that the residuals from the estimated regression relationships will be significantly serially autocorrelated. Nevertheless, trends in the regression intercepts and slopes provide a useful description of broad adjustments in the relationships between the variables throughout the ablation season (Table 3). The following generalizations can be made: *firstly*, the relationships between time series are stronger (as indicated by the coefficient of determination) for the alpine than the arctic basin; *secondly,* the intercepts and slopes of the conductivity-discharge relationships are similar although slightly steeper for the alpine in comparison with the arctic basin, and *finally* the alpine suspended sediment concentration-discharge relationships are weak in period 1, are steepest and strongest in period 2 and then stabilize at an intermediate slope for the remainder of the ablation season, whereas the arctic relationships exhibit an increasing slope as the ablation season progresses.

Many of the estimated ARIMA models in Table 4 have a "seasonal" component which reflects the impact of diurnal variations in incident solar radiation on the hydrology of both the arctic and alpine glaciers. In both arctic and alpine basins, the ARIMA models estimated for the discharge series develop a diurnal component as the ablation season progresses, and the form of the ARIMA models is very similar between the two basins. However, there are differences between the two basins. *Firstly,* the very strong short-term autoregressive components in the alpine sediment models (with AR1 and AR2 components in many cases) are not as clear in the arctic example (only an AR1 component present) and, furthermore, the diurnal component of the arctic ARIMA models is usually accommodated simply by seasonal differencing, whereas there is also a seasonal MA component for the alpine basin. *Secondly*, the electrical conductivity ARIMA models for the alpine basin virtually all have a diurnal component, but this is rarely the case for the arctic basin.

DISCUSSION AND CONCLUSIONS

From the above results, it is possible to identify the following differences between the hydrology of the arctic (Austre Brøggerbreen) and alpine (Haut Glacier d'Arolla) glacier basins:

(a) Proglacial discharge exhibits clear diurnal variations in response to solar radiation inputs in both basins. These diurnal variations can be represented by time series models of very similar form in both environments and the diurnal rhythm in the discharge record strengthens as the ablation season progresses.

(b) Suspended sediment concentrations also exhibit clear diurnal variations in both environments, but the diurnal influence (as represented by the forms of the estimated ARIMA models) is stronger in the alpine than in the arctic environment. The

Table 3 Simple regression relationships estimated between electrical conductivity/$\log_{10}$ suspended sediment concentration and $\log_{10}$ discharge at the best temporal match position.

Dependent variable (lag in h)	Independent variable	Period	a	t_a	b	t_b	n	R^2
Haut Glacier d'Arolla: 1989								
cond (0)	$\log_{10} Q$	1	108.7	42.8	−27.6	30.3	193	0.827
cond (+1)	$\log_{10} Q$	2	96.0	109.1	−23.9	86.2	139	0.982
cond (+1)	$\log_{10} Q$	3A	83.9	91.5	−20.3	71.2	245	0.954
cond (+1)	$\log_{10} Q$	3B	75.9	32.5	−17.1	24.7	236	0.722
cond (+1)	$\log_{10} Q$	4	83.3	66.5	−18.3	51.1	628	0.806
cond (+2)	$\log_{10} Q$	5	74.0	44.7	−15.7	33.8	381	0.751
$\log_{10} S$ (0)	$\log_{10} Q$	1	1.75	6.6	0.31	3.2	137	0.072
$\log_{10} S$ (0)	$\log_{10} Q$	2	−0.15	0.6	0.86	11.7	116	0.545
$\log_{10} S$ (0)	$\log_{10} Q$	3	−3.95	28.4	1.98	47.2	506	0.815
$\log_{10} S$ (−2)	$\log_{10} Q$	4	−0.87	5.2	1.09	22.9	658	0.445
$\log_{10} S$ (0)	$\log_{10} Q$	5	−1.07	7.3	1.21	29.2	490	0.443
Haut Glacier d'Arolla: 1990								
cond (0)	$\log_{10} Q$	1	90.1	160.7	−22.7	115.7	492	0.965
cond (0)	$\log_{10} Q$	2A	69.0	58.6	−15.9	44.4	196	0.910
cond (0)	$\log_{10} Q$	2B	73.5	22.8	−15.5	16.9	110	0.724
cond (+1)	$\log_{10} Q$	3	99.2	58.2	−23.0	47.5	527	0.811
cond (+1)	$\log_{10} Q$	4	85.2	43.2	−19.1	33.8	323	0.780
cond (+1)	$\log_{10} Q$	5	81.7	56.1	−18.1	42.3	290	0.861
$\log_{10} S$ (0)	$\log_{10} Q$	1	−1.16	6.8	1.10	18.6	390	0.471
$\log_{10} S$ (0)	$\log_{10} Q$	2	−5.21	30.0	2.39	46.4	310	0.875
$\log_{10} S$ (0)	$\log_{10} Q$	3	−0.27	1.9	0.95	23.0	550	0.490
$\log_{10} S$ (0)	$\log_{10} Q$	4	−0.87	3.8	1.18	18.4	288	0.541
$\log_{10} S$ (0)	$\log_{10} Q$	5	−0.27	1.1	1.07	32.8	290	0.789
Austre Brøggerbreen: 1991								
cond (0)	$\log_{10} Q$	1	no significant relationship					
cond (−2)	$\log_{10} Q$	2	84.6	33.9	−18.1	20.1	92	0.816
cond (−3)	$\log_{10} Q$	3	41.9	21.2	−7.2	11.6	515	0.208
cond (−4)	$\log_{10} Q$	4	54.5	15.0	−12.0	10.8	144	0.450
Austre Brøggerbreen: 1992								
cond (0)	$\log_{10} Q$	1	161.1	14.6	−41.9	12.7	184	0.470
cond (−2)	$\log_{10} Q$	2	26.6	33.4	−3.8	14.7	155	0.585
cond (0)	$\log_{10} Q$	3	33.6	31.4	−5.8	16.3	68	0.799
cond (−3)	$\log_{10} Q$	4	63.2	18.6	−14.6	13.8	132	0.593
cond (−2)	$\log_{10} Q$	5	57.6	17.6	−12.7	13.1	123	0.584
cond (0)	$\log_{10} Q$	6	64.4	36.7	−14.2	28.3	114	0.876
$\log_{10} S$ (0)	$\log_{10} Q$	1	−0.36	0.4	0.73	2.3	64	0.078
$\log_{10} S$ (+6)	$\log_{10} Q$	2	−0.05	0.2	0.60	6.4	53	0.444
$\log_{10} S$ (+5)	$\log_{10} Q$	3	−0.48	1.4	0.75	7.0	44	0.535
$\log_{10} S$ (+5)	$\log_{10} Q$	4	−2.89	4.2	1.47	6.9	44	0.528
$\log_{10} S$ (+2)	$\log_{10} Q$	5	−4.18	3.5	1.80	5.1	46	0.368
$\log_{10} S$ (+1)	$\log_{10} Q$	6	−3.54	3.8	1.68	6.2	44	0.476

Q: discharge in l s^{-1}; S: suspended sediment concentration in mg l^{-1}; cond: electrical conductivity in μS cm^{-1}.
a: regression intercept and t statistic (t_a); b: regression slope and t statistic (t_b).

steepest and strongest linear regression relationships between suspended sediment concentration and discharge occur relatively early in the ablation season in the alpine basin and, thereafter, lower suspended sediment concentrations are associated with given levels of discharge, suggesting a degree of exhaustion in the supply of

Table 4 ARIMA models estimated from the river time series – all models listed have parameters which are significantly different from zero ($P < 0.05$) and reduce the original time series to white noise.

Variable	Time period	ARIMA model	ARIMA parameters: 1	2	3	4	Constant (if est.)
Haut Glacier d'Arolla: 1989							
$\log_{10} Q$	1	(1,1,0)	−0.29				
	2	$(1,1,0)(0,1,1)_{24}$	0.38	0.56			
	3	$(1,1,0)(0,1,1)_{24}$	0.53	0.82			
	4	$(2,1,0)(0,1,1)_{24}$	0.24	−0.18	0.81		
	5	$(2,1,0)(0,1,1)_{24}$	0.18	−0.23	0.76		
cond	1	(1,1,0)	0.34				
	2	$(1,1,0)(0,1,1)_{24}$	0.56	0.41			
	3	$(1,1,0)(0,1,1)_{24}$	0.42	0.42			
	4	$(1,1,0)(0,1,1)_{24}$	0.12	0.73			
	5	$(0,1,0)(0,1,1)_{24}$	0.64				
$\log_{10} S$	1	(0,1,1)	0.57				
	2	(0,1,1)	0.26				
	3	$(2,0,0)(0,1,1)_{24}$	0.77	0.16	0.76		
	4	$(2,0,0)(0,1,1)_{24}$	0.67	0.24	0.85		
	5	$(2,0,0)(0,1,1)_{24}$	0.59	0.25	0.68		
Haut Glacier d'Arolla: 1990							
$\log_{10} Q$	1	$(1,1,0)(0,1,1)_{24}$	0.80	0.77			
	2	$(1,1,0)(0,1,1)_{24}$	0.46	0.81			
	3	$(1,1,0)(0,1,1)_{24}$	0.51	0.71			
	4	$(1,1,0)(0,1,1)_{24}$	0.15	0.68			
	5	$(2,1,0)(0,1,1)_{24}$	0.33	−0.35	0.48		
cond	1	$(1,1,0)(0,1,1)_{24}$	0.77	0.86			
	2	$(1,1,0)(0,1,1)_{24}$	0.41	0.83			
	3	$(2,1,0)(0,1,2)_{24}$	0.33	−0.29	0.83	−0.16	
	4	$(2,1,0)(0,1,1)_{24}$	0.05	−0.31	0.94		
	5	$(2,1,0)(0,1,1)_{24}$	0.22	−0.32	0.63		
$\log_{10} S$	1	$(2,0,0)(0,1,1)_{24}$	0.64	0.24	0.69		
	2	$(2,0,0)(0,1,1)_{24}$	0.74	0.21	0.62		
	3	$(2,0,0)(0,1,1)_{24}$	0.57	0.25	0.81		
	4	$(1,0,0)(0,1,1)_{24}$	0.62	0.74			
	5	$(2,0,0)(0,1,1)_{24}$	0.45	0.19	0.65		
Austre Brøggerbreen: 1991							
$\log_{10} Q$	1*	(1,0,0)	0.95				0.15
	2	(2,1,0)	0.42	0.38			
	3	$(1,1,0)(0,1,1)_{24}$	0.20	0.95			
	4	$(1,1,0)(0,1,1)_{24}$	0.53	0.83			
cond	1*	(1,1,0)	0.97				
	2	(2,1,0)	0.58	0.23			
	3	no satisfactory model estimated					
	4	(2,1,0)	0.61	−0.25			
Austre Brøggerbreen: 1992							
$\log_{10} Q$	1	$(0,1,0)(1,1,0)_{24}$	0.64				
	2	$(1,1,0)(0,1,1)_{24}$	0.50	0.85			
	3	$(1,1,0)(0,1,1)_{24}$	0.63	0.60			
	4	$(1,1,0)(0,1,1)_{24}$	0.42	0.76			
	5	$(1,1,0)(0,1,1)_{24}$	0.70	0.72			
	6	$(1,1,0)(0,1,1)_{24}$	0.50	0.42			
cond	1	$(0,1,0)(1,1,0)_{24}$	−0.33				
	2	(2,1,0)	0.45	0.22			
	3	(2,1,0)	0.47	0.34			
	4	(2,1,0)	0.48	0.25			
	5	(2,1,0)	0.46	0.28			
	6	(1,1,0)(0,1,1)	0.51	−0.68			
$\log_{10} S$	1	$(1,0,0)(1,1,0)_8$	0.45	−0.43			
	2	$(0,0,0)(1,1,0)_8$	0.33				
	3	$(1,0,0)(0,1,0)_8$	0.78				
	4	$(1,0,0)(0,1,0)_8$	0.58				
	5	$(1,0,0)(0,1,0)_8$	0.65				
	6	(1,0,0)	0.80				0.46

The seasonal ARIMA models relate to 24 h periods which are represented by 24 observations for all series apart from Austre Brøggerbreen 1992 $\log_{10} S$ series which has eight observations in each 24 h "season".
* ARIMA models for period 1, 1991, are estimated from short time series with many missing values.

sediment to the proglacial river (Gurnell, 1987). In contrast, the linear regression relationships steepen through the ablation season in the arctic basin. Furthermore, where there is a short or negligible lag/lead between the suspended sediment and discharge series in the alpine basin, suspended sediment concentrations lead discharge by 6 h, falling to 1 h through the 1992 observation period in the arctic basin. The steepening linear regression relationships estimated for the arctic basin confirm the apparent increase in suspended sediment availability that has been described elsewhere (Bogen, 1991) and has been attributed to the development of a subglacial drainage network (Repp, 1988). As the ablation season progresses, meltwater is generated initially from predominantly proglacial snowmelt sources and then increasingly from glacial snowmelt and icemelt sources. The decreasing lead of suspended sediment concentration in comparison with the discharge record may indicate an even more pronounced up-glacier migration of the predominant suspended sediment sources from proximal locations to increasingly distal glacial (subglacial, supraglacial and ice marginal) locations, and thus could reinforce the potentially increasing role of subglacial drainage. However, in arctic basins sediment availability is only partly governed by meltwater-sediment contact. The temperature regime of the sediment is also important and so increasing sediment availability and decreasing sediment lead times could also indicate the significance of spatially and temporally variable ground thaw processes.

(c) The linear regression relationships estimated between electrical conductivity and discharge are very similar, although the arctic relationships are not quite as steep or strong as the alpine ones for the two environments, illustrating the widely-recognized inverse relationship (Fenn, 1987). In alpine environments a variety of mechanisms have been suggested to explain this inverse relationship including the mixing of waters from different conceptual reservoirs (e.g. Brown & Tranter, 1990) and the increasing residence time of water draining from single reservoirs under recession flows (e.g. Souchez & Lemmens, 1987). The inverse relationships identified for alpine environments have been noted at a variety of timescales, with the very strong inverse pattern between electrical conductivity and discharge at the diurnal timescale (e.g. Gurnell *et al.*, in press) being emphasized as an indicator of the mixing of fast-transit, supraglacial meltwaters with longer-residence subglacial water. The analysis presented in this paper suggests that it is longer-term patterns which are most influential in the inverse relationship between electrical conductivity and discharge in the arctic basin. The ARIMA models estimated for the arctic electrical conductivity time series rarely exhibited a significant "seasonal" (diurnal) component, and the cross-correlation functions estimated between discharge and electrical conductivity did not exhibit the strong diurnal pattern of those estimated from the alpine basin time series. This suggests that the meltwater mixing processes that have received so much research attention in the alpine environment (Fenn, 1987), are less significant or operate at a different timescale in the arctic basin. Given the cold-base of much of the arctic glacier, this is hardly surprising, since the interaction between longer-residence high-solute subglacial waters (from the warm-based section of the glacier?) and shorter-residence low-solute waters will be difficult and there may be substantial time delays before the impact of the diurnal melt cycle (which is weaker than its alpine equivalent) on such mixing can be observed at a proglacial monitoring site.

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Variability in discharge, stream power, and particle-size distributions in ephemeral-stream channel systems

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Abstract The interacting factors of climate, geology, vegetation, soils, land use, and transmission losses affect the characteristics of discharge and sediment yield in ephemeral streams in arid and semiarid areas of the southwest USA. Research results are presented which describe and summarize these factors and emphasize the consequences of spatially varying rainfall and transmission losses (infiltration losses to stream bed and banks) on the subsequent spatial variability of peak discharge, stream power, and median particle-sizes of bed sediment in ephemeral-stream channels of the Walnut Gulch Experimental Watershed, Arizona, USA.

INTRODUCTION

Spatial and temporal variability in hydrologic processes and the resulting erosion and sedimentation processes are characteristically high in arid and semiarid regions. High variability results from climatic factors such as infrequent and spotty precipitation (i.e. Sellers, 1964; Osborn, 1983). Variable geologic and geomorphic features, including ephemeral-stream channels (i.e. Leopold & Miller, 1956; Thornes, 1977), and marked variations in soils and soil moisture result in variations in vegetation, land use and management (i.e. Fuller, 1975; Branson *et al.*, 1981).

Insufficient knowledge concerning spatial and temporal variations in hydrologic, erosion, and sedimentation processes and their links with geomorphic features at various scales is limiting our ability to model these processes accurately, and thus, to develop the predictive capability required for land use and management decisions. The purpose of this paper is to report the results of a hydrologic modelling study conducted to emphasize the consequences of spatially varying rainfall and transmission losses (infiltration losses to stream bed and banks) on the subsequent spatial variability of peak discharge, stream power, and median particle-sizes of bed sediment in ephemeral-stream channels of the Walnut Gulch Experimental Watershed, Arizona, USA.

DESCRIPTION OF STUDY SITES AND DATA

The Walnut Gulch Experimental Watershed, operated by the US Department of Agriculture, Agricultural Research Service (USDA-ARS) is illustrated in Fig. 1 and

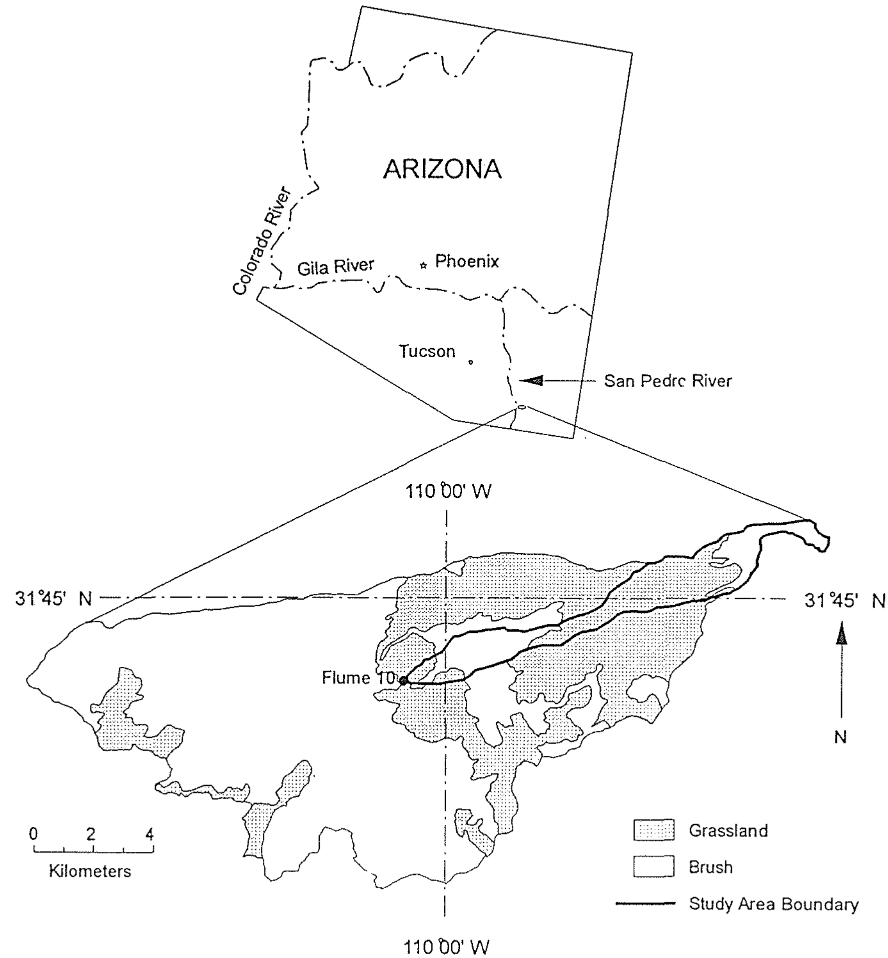

Fig. 1 Location map for the Walnut Gulch Experimental Watershed.

Subwatershed 10 is shown in Fig. 2. Subwatershed 10 drains approximately 10% of the area drained by Walnut Gulch, has relatively more relief, has a higher drainage density, and is significantly more elongated. Detailed descriptions of the Walnut Gulch Experimental Watershed, its database, and observations and research findings are given by Renard (1970) and Renard *et al.* (1993).

Mean annual temperature at Tombstone, Arizona (within the Walnut Gulch Watershed) is 17.6°C, mean annual precipitation is 324 mm, and the climate is

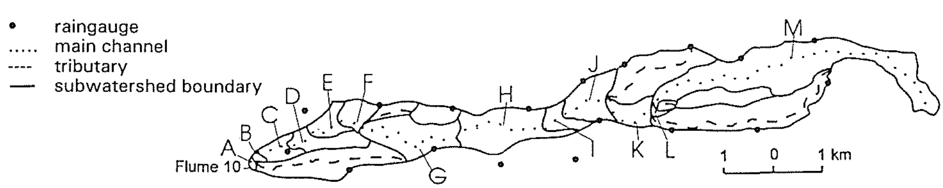

Fig. 2 Subwatershed 10 on Walnut Gulch Experimental Watershed showing channel system and subwatershed discretization for the distributed hydrologic model.

classified as semiarid or steppe. About 70% of the annual precipitation occurs during the summer months from convective thunderstorms of limited areal extent.

Soils on the Walnut Gulch Watershed, like most desert and semidesert soils, are notable for their variations with topographic features and their close relationships with the parent material because of slow rates of soil formation processes in moisture deficient environments. The parent material is dominated by fan deposits, mostly derived from intrusive and volcanic rocks and cemented with calcretes; thus, associated soils are generally well-drained, calcareous, gravelly to cobbly loams. Other important soils developed from igneous, intrusive materials and are typically shallow, cobbly, and fine textured. Finally, soils in flood plains along the ephemeral stream channels are formed of alluvium and vary from sands to loams.

Shrub vegetation, such as creosote bush, acacia, tarbush, and small mesquite trees, dominates (30 to 40% canopy cover) the lower two-thirds of the watershed. The major grass species (10 to 80% canopy cover) on the upper third of the watershed are the gramma grasses, bush muhley, and lovegrass, with some invasion of the shrub species and mesquite (Renard *et al.*, 1993). Land use consists primarily of grazing, recreation, mining and some urbanization.

METHODS AND ANALYSES

Distributed watershed modelling

A calibrated, distributed hydrologic model (Lane, 1982) was used as a tool to compute runoff from rainfall data and to route the runoff in ephemeral-stream channels to compute peak discharge and stream power. Spatial variations in peak discharge due to distributed rainfall, soils, vegetation, and transmission losses are explicitly included in the calculations.

Thiessen weights were determined for the 18 recording raingauges on or near Subwatershed 10 (Fig. 2) and then areal average rainfall was determined for each of the 38 upland and lateral flow areas used to represent the subwatershed. This procedure was repeated for 74 individual runoff producing storms over the 11-year period of record from 1967 to 1977 to fit, or calibrate, the model to observed runoff data measured in the supercritical flume (Fl 10) located at the subwatershed outlet.

This fitting procedure constituted the model calibration with the following results for the 74 runoff events:

$$V_f = 0.42 + 0.89V_o \qquad (1)$$

with a value of $R^2 = 0.71$ where V_f is the fitted runoff volume (mm) and V_o is the observed runoff volume (mm). The corresponding equation for peak discharge is:

$$q_f = 0.51 + 0.96q_o \qquad (2)$$

with a value of $R^2 = 0.73$ where q_f is the fitted peak discharge (mm h^{-1}) and q_o is the observed peak discharge (mm h^{-1}).

On 9 July 1993 a thunderstorm occurred over the upper portion of Subwatershed 10 and produced runoff at the subwatershed outlet. Runoff curve numbers were adjusted for the dry initial condition until the model estimate matched the runoff peak discharge as measured at the flume. Stream channel cross sections and composite bed material samples were obtained before and after this runoff event at each cross section.

Finally, 60-minute point rainfall amounts for the 2 and 10 year return periods were determined following Osborn (1983) and then adjusted using a depth area relationship (Osborn, 1983) to estimate average rainfall depths over the 16.6 km^2 subwatershed. These subwatershed-average rainfall amounts were used as input to the calibrated, distributed model to produce runoff volume and peak discharge estimates for the 2 and 10 year floods.

Stream power and sediment transport

Stream power per unit length of the stream bed is calculated as:

$$P = \gamma QS \tag{3}$$

where P is stream power in N s^{-1}, γ is the specific weight of water (N m^{-3}), Q is the discharge rate (m^3 s^{-1}), and S is the longitudinal slope of the channel bed. Stream power has been related to total sediment transport (Bagnold, 1960, 1966, 1977). Stream power per unit weight of water, called unit stream power, has been related to total sediment concentration in streams (i.e. Yang & Stall, 1976; Yang & Molinas, 1982). Graf (1983) used stream power per unit length as a surrogate for total sediment transport in ephemeral stream channels.

Earlier, Lane (1955) recognized the role of stream power in stating a qualitative relationship for stable alluvial channels. Lane's equation stated that:

$$G_s d_s \text{ is proportional to } QS \tag{4}$$

where G_s is sediment transport rate (kg s^{-1}), d_s is a characteristic sediment particle size (mm), Q is discharge rate and S is slope of the stream bed, as in equation (3). Without loss of generality, the right hand side of equation (4) can be multiplied by gamma, γ, and both sides of the equation can be divided by d_s (since both γ and d_s are positive quantities) to produce

$$G_s \text{ is proportional to } \gamma QS/d_s = P/d_s \tag{5}$$

which again suggests that stream power might be a useful surrogate for sediment transport rate.

RESULTS AND DISCUSSION

Physical characteristics of the main channel of Subwatershed 10 are summarized in Table 1. Composite bed material samples were collected at 11 cross sections (Table 1). Median particle size varied with distance along the main channel and also with time before and after the runoff event of 9 July 1993. However, there were no statistically significant trends with distance along the main channel and no statistically significant

Table 1 Physical characteristics for the main channel of Watershed 10 as measured in the field and on 1:5000 scale ortho-topographic maps. Channel characteristics used in the distributed hydrologic model to simulate runoff.

Channel reach	Reach length (km)	Distance above Fl 10 at lower end of reach (km)	Average width of reach (m)	Slope at lower end of reach	Median particle size: Before* (mm)	After (mm)
A) 56†-x1‡	0.18	0	9.1	0.0106	1.48	0.78
B) 50	0.21	0.18	24	0.0098	----	----
C) 47-x3	0.84	0.39	24	0.0089	2.28	0.96
D) 44-x4	0.68	1.22	23	0.0163	1.45	1.71
E) 41-x5	0.9	1.9	17	0.0157	1.72	1.94
F) 38	0.6	2.8	18	0.014	----	----
G) 31-x6	3.19	3.39	12	0.0124	0.95	0.76
H) 28-x7	2.27	6.58	14	0.0111	1.38	2.03
I) 25-x8	0.58	8.85	18	0.0131	1.89	1.31
J) 22-x9	1.06	9.43	20	0.0097	1.41	1.23
K) 16-x11	1.5	10.49	12	0.0105	2.17	0.96
L) 13-x14	0.18	11.99	7.6	0.0127	1.28	0.96
M) 03-x13	7.42	12.16	9.1	0.0114	2.98	1.37
Upper end	----	19.58	----	----	----	----

* Samples taken before and after the first runoff event of the season on 9 July 1993.
† Channel reach numbers as represented in the distributed model.
‡ Denotes cross section numbers on main channel where bed material samples were taken.

differences in median particle sizes before and after the runoff event of 9 July 1993.

Hydrologic variable estimates based on application of the calibrated, distributed hydrologic model are summarized in Table 2. Calculated peak discharge rates along the main channel in Subwatershed 10 for the storm of 9 July 1993, and for the 2 and 10 year floods are shown in Fig. 3(a). Corresponding stream power results are shown in Fig. 3(b).

Excluding the boundary point at the upper end of the main channel, the ratio of maximum to minimum values for the channel characteristics in Table 1 varied by a factor of approximately 2 to 3. The corresponding maximum to minimum ratio for peak discharge of the 9 July 1993 storm is 5.5 and for stream power is 6.4. Ratios for peak discharge of both the 2 and 10 year floods are about 1.6 and ratios for stream power are 2.0. Recall that rainfall input to the model for the storm on 9 July 1993 was distributed over the 38 elements used to model Subwatershed 10 (Fig. 2) while the rainfall input for the 2 and 10 year floods was calculated from a depth area relation and thus was assumed to be uniform over the entire subwatershed. These analyses suggest that the assumption of uniform rainfall input to the hydrologic model significantly underestimated the spatial variability of peak discharge and stream power, and thus by inference, erosion and sediment transport rates.

The results presented in Table 2 are based on modelling results after the model was calibrated using observed runoff data measured at the subwatershed outlet. However, peak discharge and stream power values calculated at interior points remain unvalidated.

Table 2 Hydrologic variable estimates for the main channel of Watershed 10 based on the physical characteristics shown in Table 1 and results of applying the distributed hydrologic model. Calculations are for the storm of 9 July 1993 and for the 2 and 10 year floods.

Channel reach	Distance above Fl 10 at lower end of reach (km)	9 July 1993: Q* ($m^3 s^{-1}$)	P† ($N s^{-1}$)	2 year: Q ($m^3 s^{-1}$)	P ($N s^{-1}$)	10 year: Q ($m^3 s^{-1}$)	P ($N s^{-1}$)
A) 56‡-x1§	0	4.79	498	14.6	1520	44.5	4620
B) 50	0.18	5.01	480	12.2	1170	37.5	3600
C) 47-x3	0.39	5.44	476	12.6	1100	38.5	3350
D) 44-x4	1.22	7.39	1180	13.9	2220	41.8	6680
E) 41-x5	1.9	8.84	1360	14.4	2220	42.9	6610
F) 38	2.8	10.5	1430	15.3	2100	45.2	6210
G) 31-x6	3.39	11.9	1450	15.1	1830	44.6	5420
H) 28-x7	6.58	18.2	1980	16.9	1840	48.7	5300
I) 25-x8	8.85	23.9	3060	17.1	2200	48.5	6220
J) 22-x9	9.43	26.2	2490	17.5	1670	49.4	4690
K) 16-x11	10.49	24.7	2540	14.6	1510	40.6	4180
L) 13-x14	11.99	15.9	1980	11.0	1370	29.5	3680
M) 03-x13	12.16	13.8	1540	11.0	1230	29.6	3310
Upper end	19.58	0	0	0.0	0	0.0	0

* Q is estimated peak discharge using the calibrated, distributed hydrologic model.
† P is stream power calculated from the estimated peak discharge.
‡ Channel reach numbers as represented in the distributed model.
§ Denotes cross section numbers on main channel where bed material samples were taken.

Adequate study of spatial variability of hydrological processes and sedimentation processes in ephemeral-stream channel systems will require continuous monitoring of discharge, hydraulic variables, and sediment concentration during runoff events, as well as monitoring of physical features of the channel systems between events at a sufficient number of interior points to test the validity of distributed modelling results.

Subwatershed 10 was discretized for modelling purposes as shown in Fig. 2. This resulted in 13 channel reaches along the main channel. The mean reach length is 1.5 km and the range of lengths is 0.18 to 7.42 km. From Fig. 3, it is apparent that at least one additional cross section (and thus subwatershed in the model discretization) is needed between the cross section at 12.16 km above the flume and the main channel headwaters at 19.58 km.

Under the special circumstances of this study, an appropriate distance between monitoring points along the main channel appears to be 1-2 km. For a watershed of this scale (main channel length of about 20 km), 10 to 20 interior measurement points are needed to test the validity of distributed hydrologic models of the complexity used in this study.

Similar studies on other subwatersheds of Walnut Gulch over a range of geomorphic features are needed to generalize these results basinwide. Such generalizations are

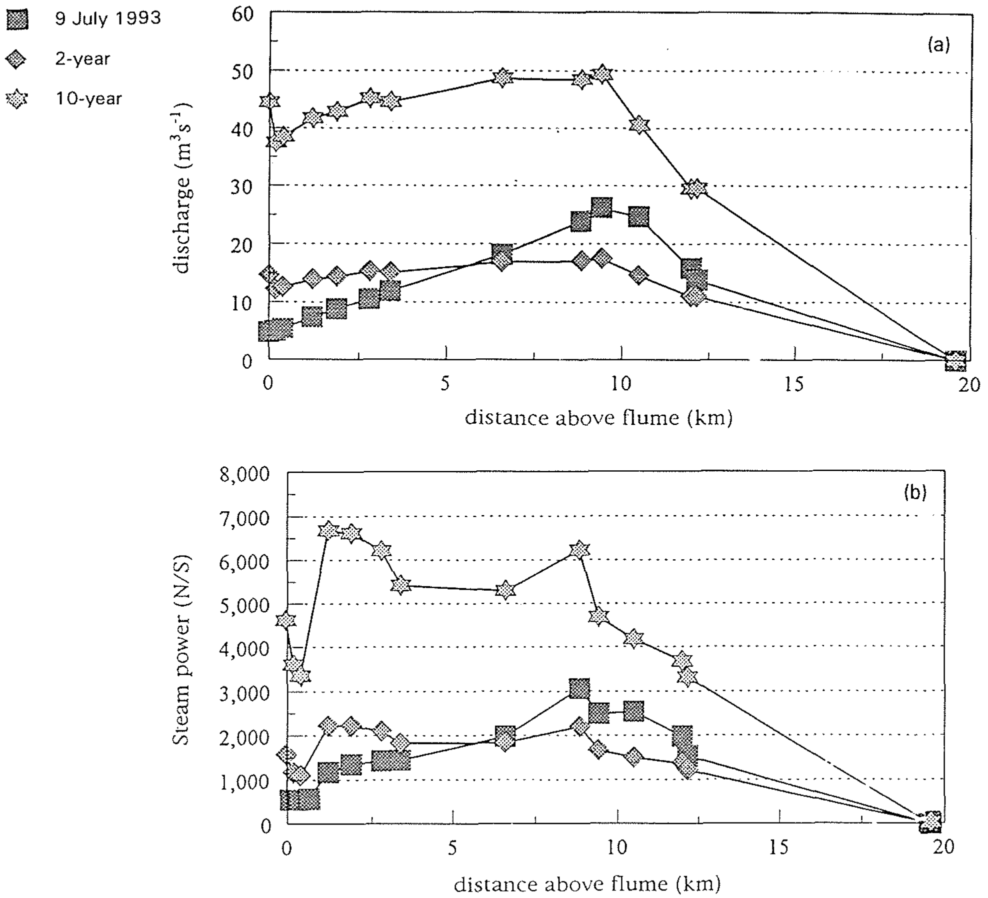

Fig. 3 Variation with distance in the main channel of Subwatershed 10 of (a) peak discharge and (b) stream power.

needed before the impacts of spatial variability in hydrologic and sedimentation processes can be understood, modelled, and predicted.

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Variability in Stream Erosion and Sediment Transport (Proceedings of the Canberra Symposium, December 1994). IAHS Publ. no. 224, 1994.

Recent changes in rates of suspended sediment transport in the Jökulsá á Sólheimasandi glacial river, southern Iceland

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Abstract This paper examines change over the 1973-1992 period in suspended sediment concentrations and loads for the Jökulsá á Sólheimasandi glacial river in southern Iceland. The main data set used is derived from all-year-round depth-integrated suspended sediment samples taken by the Icelandic National Energy Authority (Orkustofnun) from a bridge site 4 km downstream from the glacier snout. The Jökulsá á Sólheimasandi system delivers very high suspended sediment yields (mean annual value, 1973-1988, is 8990 t km^{-2} $year^{-1}$). Between 1973 and 1992, however, the trend shows that average sediment loads have decreased by 48%, from 14.6 kg s^{-1} to 7.6 kg s^{-1}, and average suspended sediment concentrations have dropped by 33% from 725 mg l^{-1} to 487 mg l^{-1}. Although *annual* runoff appears relatively stable, *melt-season* flows have become more dominant while spring and autumn runoff has declined, as have spring and autumn suspended sediment concentrations and loads. This growing flow seasonality appears to be driven by a significant warming in summer air temperatures in southern Iceland (at 0.27°C per decade) over the last twenty years. Sediment load decreases may also relate to fine-sediment exhaustion effects, possibly associated with a depletion of proglacial sediment supplies in response to substantial readvance of the basin glacier, Sólheimajökull.

INTRODUCTION

Studies of glaciofluvial sediment transport are receiving renewed attention recently, largely because of its importance in: (a) shedding light on glacial, geomorphological and hydrological processes upstream; (b) estimating sediment yields and erosion rates in inhospitable terrains; (c) influencing the quality of turbidity-sensitive aquatic habitats; (d) sediment-associated contaminant transfer; (e) water potability issues; (f) hydroelectric turbine operation; and (g) downstream siltation impacts, including delivery of sediment to nearshore zones which is often vital to the stability of coastlines. However, few data exist from subarctic environments. Recent work by the Icelandic National Energy Authority (Orkustofnun) and the author has shown that many of the glacierized basins in southern Iceland deliver very high sediment yields by world standards (in the order of 10 000 t km^{-2} $year^{-1}$) (Tómasson, 1991; Lawler *et al.*, 1992). Little is known, however, of the longer-term stability of sediment concentrations and yields in response to environmental and climatic changes at different timescales. The aim of this paper, therefore, is to make a preliminary examination of changes in suspended sediment fluxes from the Jökulsá á Sólheimasandi basin over the 1973-1992 period, in relation to climatic, hydrological and glaciological fluctuations. As far as the author is aware, this is the first all-year-round study of longer-term sediment transport trends in relation to climatic and glaciological change within a glacierized basin. Results of complementary studies of melt-season sediment transport dynamics will be published elsewhere.

RESEARCH AREA

All field measurements reported here were carried out at a natural-section gauging station below a road bridge over the Jökulsá á Sólheimasandi glacial river, the meltwater outlet of the valley glacier Sólheimajökull in southern Iceland (19°25′W; 63°30′N). The station is 4 km from the glacier snout – the first point at which all braided meltwater rivers are collected together in a single-thread reach. The sampling site is the nearest to a glacier snout of any of those operated by Orkustofnun. Drainage area at the bridge is around 110 km^2, of which approximately 71% is ice-covered, and bankfull discharge is estimated to be 100 $m^3 s^{-1}$ (Lawler, 1991; Lawler *et al.*, 1992). Annual precipitation (1931-1960) is estimated to rise from at least 1600 mm at the bridge gauging site (altitude 50 m) to over 4 000 mm near the highest boundary of the catchment (1493 m) (Eythorsson & Sigtryggsson, 1971 (cited in Björnsson, 1979)). Hydroclastic and acid volcanic rocks dominate the basin, and palagonite tuffs and breccias are common (Carswell, 1983). Anthropogenic impact in the uninhabited and undeveloped Jökulsá á Sólheimasandi basin is minimal. There is no evidence that basin land use has changed during the study period. This makes it ideal for the study of the impact of natural environmental forcing on sediment transport. A full description of the study area can be found in Lawler (1991).

Mean annual suspended sediment yield (1973-1988) is 8990 t km^{-2} $year^{-1}$ (standard error = 1285 t km^{-2} $year^{-1}$ or 14.3%). Likely suspended sediment sources include: supraglacial material in the ablation zone of the glacier tongue (Lawler, 1994, his Fig. 1); voluminous bands of englacial debris, including tephra; subglacial sediments; proglacial moraines, terraces, channel and sandur deposits; and extra-glacial hillslopes which are deeply gullied and sparsely vegetated.

DATA SOURCES

Continuous flow and sediment records are rarely available for the more remote high-latitude regions of the world. The main Orkustofnun data set used in this paper comprises, for the 1973-1992 period, the 226 values for mean instantaneous suspended sediment concentration (bulked from around 1200 individual samples), for which discharge measurements were available. Field sampling techniques and laboratory methods are discussed by Lawler (1991) and Lawler *et al.* (1992).

The main strengths of the Orkustofnun dataset are: use of recognized samplers (e.g. US D-49)); full depth- and width-integrated sampling; *measurement* of discharge at the time of sampling (not simple *estimation* from rating curves subject to substantial error associated with rapidly-changing channel geometries (Lawler *et al.*, 1992); a reasonably long data series, beginning in 1963 (1973 for the Jökulsá á Sólheimasandi); all-year-round sampling; and representation of a range of subarctic environments, which are relatively under-researched in global terms. Disadvantages, however, include a low sampling frequency (approximately monthly) which is not designed to detect transient and diurnal sediment pulses (e.g. Lawler & Brown, 1992; Lawler *et al.*, 1992), and lack of continuous river flow data at a number of sampling stations, including the Jökulsá á Sólheimasandi.

CHANGES IN SUSPENDED SEDIMENT LOAD

Figure 1 confirms the very high sediment loads (calculated as the products of the instantaneous suspended sediment concentration and discharge values) characteristic of this system (Tómasson, 1991; Lawler *et al.*, 1992). Although much of the scatter in Fig. 1 is seasonal variation normally associated with the regularities of melt seasons in glaciofluvial systems, such variability can hinder the identification of trend in the data. Nevertheless, the trend in Fig. 1 shows a statistically significant decline in average sediment loads of 48% between 1973 and 1992, from 14.6 kg s^{-1} to 7.6 kg s^{-1}. Note also how the annual minima decline at a faster rate than the maxima (Fig. 1): the system today, then, is still apparently able to generate high loads, but instances of low sediment flux are now much more common. Many possible causes of sediment flux decline can be hypothesized, related to changes in sediment production mechanisms, source locations, supply volumes, access processes, delivery pathways, transport rates and storage opportunities, and these will be addressed elsewhere. Space permits here simply a brief exploration of covariance in some relevant hydrological and glaciological variables, which possibly point to shifts in the sediment supply and transport regimes.

There is no evidence from these "snapshot" discharge measurements to suggest that *annual* runoff has declined during the study period, as has been found by Snorrason (1990) elsewhere in Iceland. Instead, it seems that significant suspended sediment

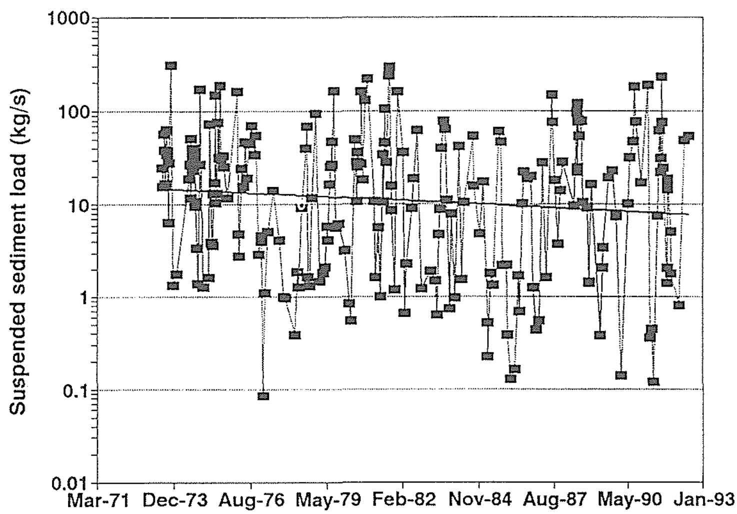

Fig. 1 Changes in suspended sediment load (SSL; kg s^{-1}) for the Jökulsá á Sólheimasandi, 1973-1992. The fitted linear regression, significant at $p < 0.05$, is: log SSL = 2.243893 − 0.0000403 D, where D is number of days from 1 January 1900.

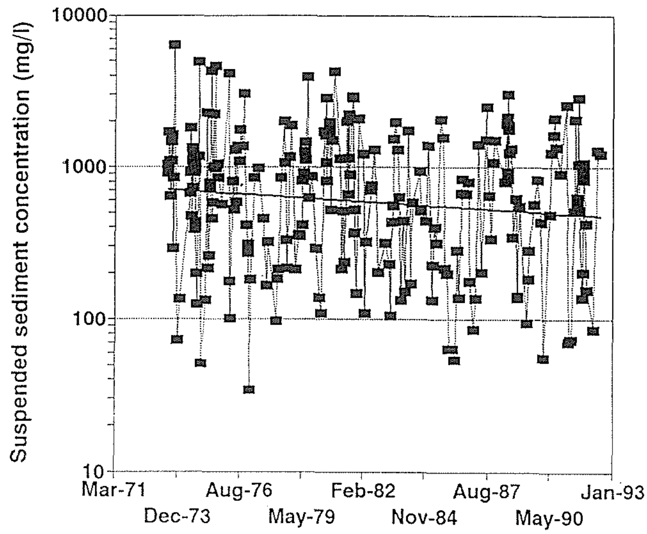

Fig. 2 Changes in suspended sediment concentration (SSC; mg l^{-1}) for the Jökulsá á Sólheimasandi, 1973-1992. The fitted linear regression, significant at $p < 0.05$, is: $\log SSC = 3.522651 - 0.0000247\,D$, where D is number of days from 1 January 1900.

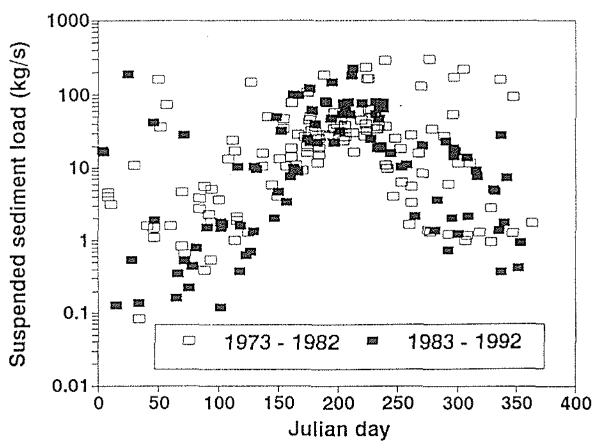

Fig. 3 Seasonal variation in suspended sediment load for each decade of record, 1973-1982 and 1983-1992. Note the greater number of lower values in the spring and autumn seasons of the more recent period, but a preserved peak around day 210.

concentration decreases are important here, the regression trend suggesting a drop of 33% from 725 mg l^{-1} in 1973 to 487 mg l^{-1} in 1992 (Fig. 2). Furthermore, although the *summer* peak in sediment loads has been sustained in the later decade, in *spring* (Julian day 60-150) lower values are now more common and, in *autumn* (day 250-330), the higher values are lacking (Fig. 3). This pattern is mirrored by the suspended sediment concentration and discharge values which are positively correlated (e.g. Lawler, 1991). Sampled spring and autumn river flows (and presumably transportational energy) tend now to be lower, while summer flows have been significantly enhanced ($p < 0.05$; Lawler, in preparation) (Fig. 4). This shift in flow seasonality may also have increased

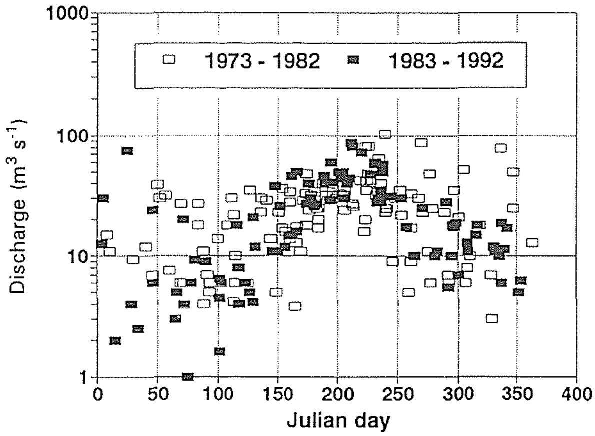

Fig. 4 Seasonal variation in discharge for each decade of record, 1973-1982 and 1983-1992. Note the greater number of lower values in the spring and autumn seasons of the more recent period, compensated for by a significant rise in summer flows.

the stability of the subglacial drainage network, with a corresponding reduction in the frequency with which new subglacial sediment source areas are "swept" before and after the main melt season.

This apparent increase in the dominance of summer flows may in turn be responding to strongly significant summer warming here since 1973 (Fig. 5). The rate of temperature increase here (0.27°C per decade) is mirrored in other arctic and subarctic environments (e.g. Farmer, 1989 (cited in Kullman, 1992); Nordli, 1991; Alexandersson & Dahlström, 1992; Jónsson, 1992) and is not inconsistent with temperature rises forecast by GCMs for similar latitudes (e.g. Carter *et al.*, 1991). The causes and projected continuity of such warming are separate issues of course.

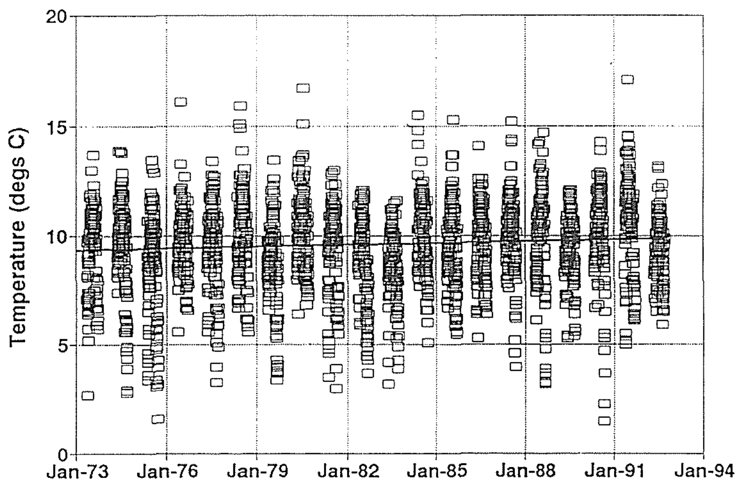

Fig. 5 Changes in mean daily summer (June-September inclusive) air temperature (T_s; °C) at Vík, on the southern coast of Iceland (21 km southeast of the gauging site), showing a distinct warming trend. The fitted linear regression, significant at $p < 0.0001$, is: $T_s = 7.35 + 0.000075\,D$, where D is number of days from 1 January 1900. Temperature rise is 0.27°C/decade. The scatter is mainly seasonal variation.

Sediment supply limitations may also be important. Scrutiny of the residuals from the discharge-suspended sediment concentration rating curve indicates that, *for a given flow*, less sediment tends to be transported today than in the 1970s (Lawler, in preparation). These lower sediment fluxes, perhaps paradoxically, coincide with an extremely active phase of glacial advance (Figs 1, 6). There may, of course, be time lags in the system whereby the smaller sediment loads of the 1980s are merely responding to earlier, relatively quiescent, glaciological phases (cf. Harbor & Warburton, 1993): if so, yields may rise again over the coming years in response to the vigorous glacial advance of the mid-1980s. Alternatively, it could be argued that higher sediment loads in the 1970s relate to peak exposure of proglacial sediment stores at that time – shortly after maximum deglaciation was achieved (Fig. 6). These stores are known to be important in some glaciofluvial contexts (e.g. Hasholt & Walling, 1992). Such sediment supplies could have been progressively depleted by rainwash and overbank flow events (if not continually renewed at comparable rates (cf. Richards, 1984)), allowing partial exhaustion phenomena to emerge in the data of later years.

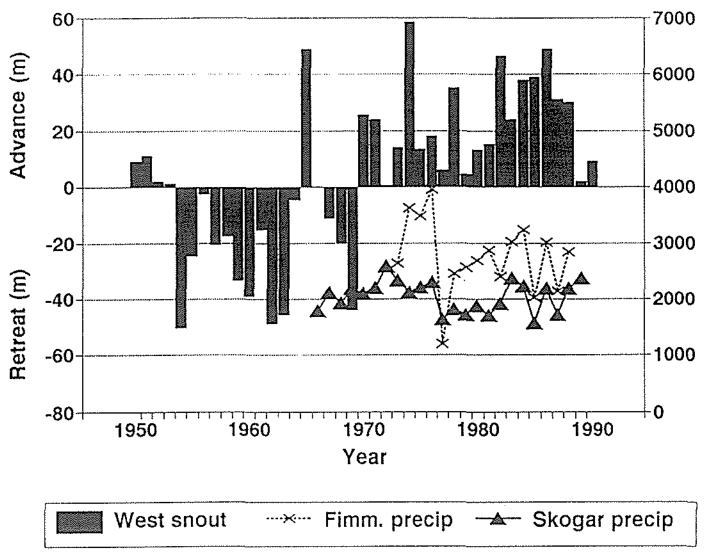

Fig. 6 Changes in the annual position of the west snout of the basin glacier, Sólheimajökull, 1950-1990 (data from *Jökull*, and Oddur Sigurdsson, personal communication). A period of uninterrupted readvance began in 1970. Precipitation at Skógar (6 km west of the gauging station) and Fimmvörduháls (Fimm.), ~2 km from the western edge of the basin at 630 m, is added for comparison in the latter part of the period.

Similar impacts at longer timescales were hypothesized within the "paraglaciation" concept of Church & Ryder (1972). Such complex responses, with so many possible reinforcing, cancelling and inertial effects associated with glacier motion and snout advance/retreat cycles, underline the need for extreme caution in making overly-simple extrapolations about the likely impact of longer-term glaciation on sediment yields and erosion rates (see recent debates involving Hicks *et al.*, 1990; Molnar & England, 1990; Summerfield & Kirkbride, 1992; Harbor & Warburton, 1993).

CONCLUSIONS

The following preliminary conclusions can be drawn:

(a) The trend suggests that average suspended sediment loads in the Jökulsá á Sólheimasandi system have almost halved over the 1973-1992 period.

(b) Average suspended sediment concentrations have also decreased by 33% over the same period.

(c) These statistically significant declines in suspended sediment fluxes are more noticeable in spring and autumn, and the relative stability of summer sediment transport patterns underscores the need for all-year-round and seasonally-sensitive investigations (rather than melt-season studies only) when trying to monitor the impact of environmental change in glacierized basins characterized by significant runoff in all seasons.

(d) Although no trend is identifiable in *annual* runoff, summer river discharges, as sampled, appear to have become more dominant – at the expense of spring and autumn flows. This loss of runoff volume and transportational capacity either side of the main melt season may partly explain the seasonal pattern of suspended sediment declines.

(e) Significant summer warming over the last 20 years in southern Iceland (Vík) may be driving the recent increases in flow seasonality which are helping to sustain melt-season suspended sediment export rates.

(f) Lower rates of suspended sediment transport in the 1980s are synchronous with a phase of especially active glacier snout advance: declines may be lagging behind earlier, less dynamic, glacial phases or relate to a progressive exhaustion of (proglacial) sediment supplies.

Such statistically significant changes (and year-to-year variability) in sediment load reinforce the need to base mean sediment yield estimates on a sufficiently long period of record. The magnitude of change identified here also serves to remind that, when making spatial comparisons of sediment yield between basins, allowances should be made if different periods of record are used relative to local, regional or global gradients of environmental change. Further research is in progress to define all facets of sediment transport change, including seasonal signatures; to model the relationship of change to the driving climatic, hydrological and glaciological processes and to the dynamics of melt-season suspended sediment fluxes; and to determine the spatial coherence of identified patterns across other Icelandic and subarctic systems.

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Comparative modelling of large watershed responses between Walnut Gulch, Arizona, USA, and Matape, Sonora, Mexico

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Abstract Data collected from 1967 to 1977 on the intensively gauged Walnut Gulch Experimental Watershed near Tombstone Arizona are used to calibrate the Simulator for Water Resources on Rural Basins – Water Quality (SWRRBWQ) model. The Walnut Gulch data base includes topography, soils, vegetation and land use characteristics, as well as measured precipitation and runoff data. Results of the SWRRBWQ modelling effort are extended to the upper subarea of the Matape watershed in Sonora, Mexico to evaluate the practical aspects of extending natural resource models validated on small, data rich watersheds to larger watersheds with sparse data where comparably intense validation is impossible. Available data characterizing the Matape watershed include water yield, stream channel network, and vegetation, land use, and soils characteristics developed from remotely sensed data. Model calibration results from a small, intensively gauged watershed, and modelling limitations on larger, complex watersheds are discussed.

INTRODUCTION

The use, conservation, and preservation of rangelands and marginal dryland farming areas are constrained in part by lack of knowledge and technology bases that come from continued, long-term research and technology development and transfer. By evaluating the practical aspects of extending natural resource models validated on small, data-rich watersheds to larger watersheds with sparse data where comparably intense validation is impossible, progress can be made in developing knowledge and technology to assist land managers in selecting the best from among alternative management practices.

Hydrologic response is affected by vegetation, soils, climate, and management activities. Although implementing, evaluating, and monitoring alternative management practices is expensive and time consuming, the hydrologic response of various management practices can be modelled inexpensively and quickly. In addition, modelling efforts can expose data collection needs as well as future research needs.

The objective of this research is to calibrate the Simulator for Water Resources in Rural Basins – Water Quality (SWRRBWQ) (Arnold *et al.*, 1990) model using data collected on Walnut Gulch to simulate hydrologic and related processes on the watershed. Results of the SWRRBWQ modelling effort are extended to the upper subarea of the sparsely gauged Matape watershed in Sonora, Mexico.

STUDY SITES

Walnut Gulch Experimental Watershed

The 150 km^2 Walnut Gulch Experimental Watershed in southeastern Arizona is representative of approximately 60 million hectares of rangeland found throughout the semiarid Southwest. Precipitation data have been collected on the recording raingauge network on Walnut Gulch since its completion in 1961. These data have been used to evaluate the spatial and temporal properties of precipitation. In addition to precipitation, the Walnut Gulch data base includes topography, soils, vegetation and land use characteristics, as well as runoff data. Additional information on the Walnut Gulch Experimental Watershed, its data base, and observations and research findings are given by Renard *et al.* (1993).

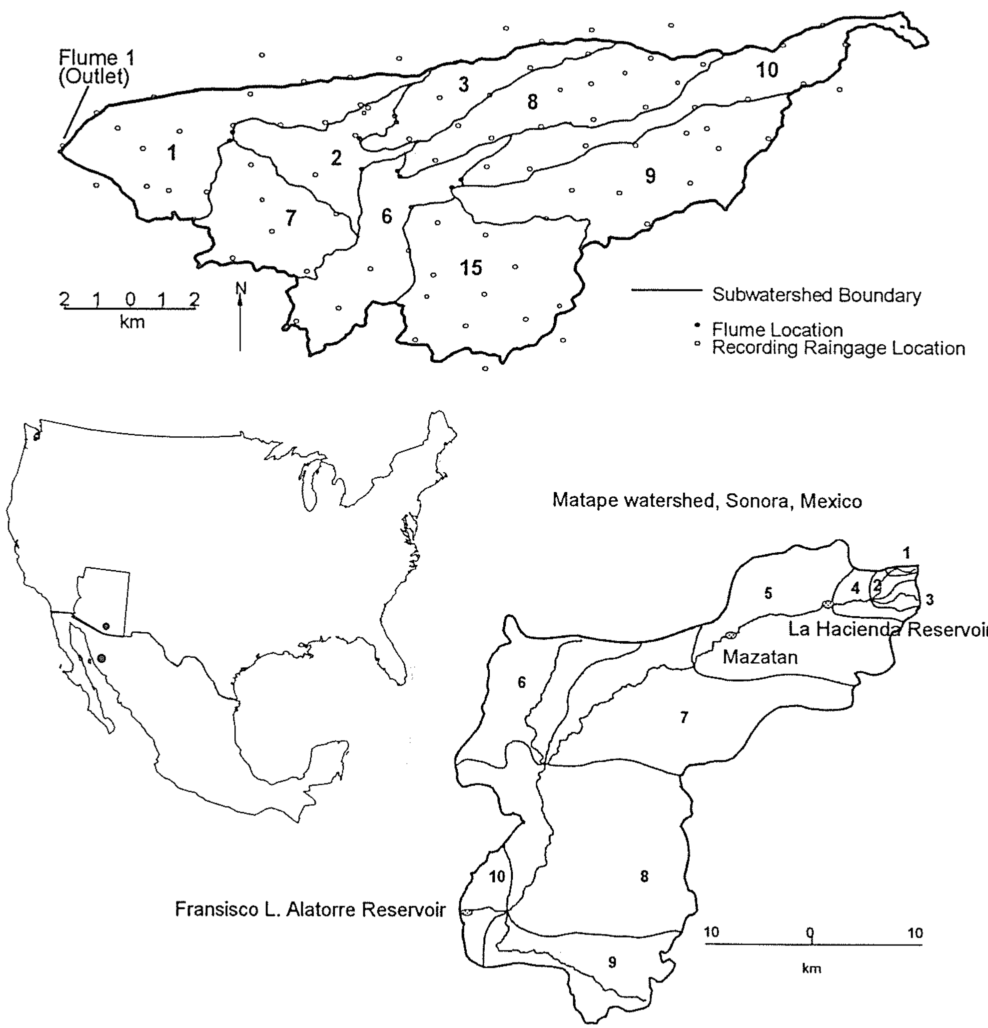

Fig. 1 Location map showing USDA-ARS Walnut Gulch Experimental Watershed, Arizona, USA, and Matape watershed, Sonora, Mexico, subwatershed discretization for SWRRBWQ modelling.

Upper Matape watershed

The 3140 km^2 Rio Matape watershed is typical of millions of hectares of Lower Sonoran desert and grasslands (Brown, 1982) in northern Mexico. The watershed drains into the Fransisco L. Alatorre Reservoir. Elevation of the watershed ranges from approximately 240 m above mean sea level at the reservoir to about 1680 m at the headwaters, about 34 km northeast of Mazatan. Available data for model calibration consists of limited streamflow records and sparse precipitation data. Subwatershed and stream channel network characteristics were developed from remotely sensed data and information digitized in a geographic information system (GIS).

METHODS AND ANALYSES

Distributed watershed modelling with SWRRBWQ

SWRRBWQ is a long-term water and sediment yield simulator that was developed for simulating hydrologic responses in rural basins. The model can be used to predict the effect of management decisions on water and sediment yields, as well as water quality, for ungauged rural basins. Large, complex watersheds can be modelled with SWRRBWQ because large basins can be subdivided based on soils, land use, or management and each subwatershed can be modelled with different precipitation, temperature, soils and management input values. The three major components of the model are weather, hydrology, and sedimentation. Processes considered in the model include surface runoff, return flow, percolation, evapotranspiration, transmission losses, pond and reservoir storage, sedimentation, and crop growth. Detailed descriptions of these processes can be found in Arnold *et al.* (1990).

Model input values – Walnut Gulch Experimental Watershed

Watershed characteristics The Walnut Gulch watershed was divided into nine subwatersheds (Fig. 1). A supercritical measuring flume (Smith *et al.*, 1981) is located at the outlet of each subwatershed, allowing for comparison of observed and simulated water yield at interior points (subwatersheds 3, 7, 8, 9, 10 and 15), and thus model calibration. Subwatershed sizes, channel lengths, average land slope, and length of overland flow were determined from 1:24 000 scale topographic maps. Soil parameters for each subwatershed were based on a soil survey of Walnut Gulch (Gelderman, 1970). Effective hydraulic conductivity of channel alluvium was estimated as 25 mm h^{-1}. Return flow travel time was set high to minimize calculated subsurface flow. Initial runoff curve numbers (CN2) for each subwatershed were estimated based on soil groups and were adjusted during model calibration. Subwatershed topographic characteristics are summarized in Table 1.

Weather Precipitation, air temperature, and solar radiation values are used in the model. Daily precipitation data were Theissen weighted for each subwatershed to create a single, aerially averaged daily precipitation record for each subwatershed. Daily maximum and minimum air temperatures are simulated by the model using normal

Table 1 Walnut Gulch Experimental Watershed subwatershed characteristics.

Subwatershed	Size (ha)	Main channel length (km)	Average land slope
1	2210	6.76	0.020
2	963	4.12	0.020
3	898	7.24	0.022
6	1545	3.86	0.020
7	1352	5.92	0.041
8	1550	9.48	0.021
9	2359	14.45	0.027
10	1663	18.29	0.038
15	2393	4.91	0.034

distribution equations and the daily mean and standard deviation of maximum and minimum temperature for each month. Solar radiation is also simulated by the model using normal distribution equations and the daily mean and standard deviation of daily solar radiation for each month. General weather data including the TP-40 10 year frequency rainfall amounts for 0.5 h and 6 h were taken from the Rainfall Frequency Atlas of the United States (Hershfield, 1961).

Vegetation and land use The rangelands at Walnut Gulch are primarily used for domestic animal grazing and were modelled as perennial crops with no tillage. The lower portion of the watershed is dominated by shrubs and the upper portion of the watershed is grassland with some shrub invasion.

SWRRBWQ model calibration using data from Walnut Gulch Experimental Watershed

The model calibration procedure consisted of varying input values including Condition 2 curve number, soil properties, and return flow travel time within reasonable ranges of uncertainty and comparing observed and simulated mean annual water yields. Subwatersheds 3, 7, 8, 9, 10 and 15 were individually calibrated for the eleven year period 1967-1977 (Fig. 2). To compute basin outflow, the SWRRBWQ model sums subwatershed outflows. In addition to water yield, observed and simulated maximum peak flow were compared (Table 3). Basin peak flow rates are computed in the model by a modified rational equation. Although it is possible to accurately simulate water yield, the observed and simulated peak flow values do not agree, with significant underprediction in all cases (Fig. 3).

Runoff calculations in the SWRRBWQ model are based on daily precipitation. Because of the short duration of storms characteristic of Walnut Gulch, the peak flow estimation procedures in SWRRBWQ need further evaluation and may need additional modification to accurately model peak runoff. Because subwatershed peak rates are used in the model to estimate subwatershed sediment yields, no attempt was made to model sediment yields.

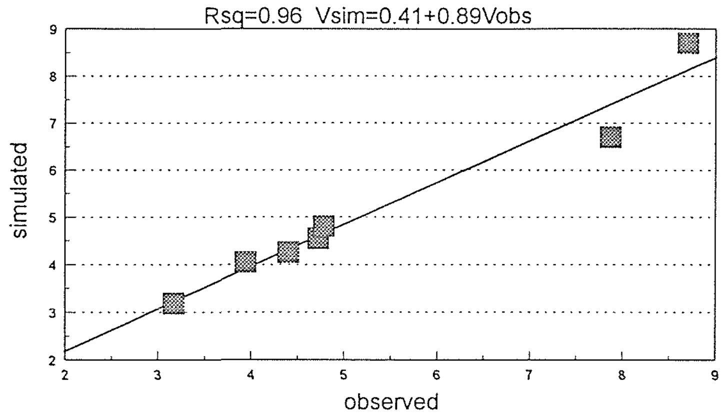

Fig. 2 Walnut Gulch Watershed observed and SWRRBWQ simulated average annual runoff (mm) 1967-1977, subwatersheds 1, 3, 7, 8, 9, 10 and 15.

Results of model simulation on the Upper Matape watershed

The Matape watershed was divided into 10 subwatersheds based on topography and soil type (Fig. 1). Subwatershed topographic and stream channel characteristics were determined using a GIS that incorporates data from 1:50 000 and 1:250 000 scale maps. Subwatershed topographic characteristics are summarized in Table 2.

Limited measured precipitation data are available for three locations on the Matape watershed. Average annual precipitation from 1980-1992 at La Hacienda Reservoir was 662 mm. Average annual precipitation at the Fransisco L. Alatorre Reservoir is approximately 390 mm. At Mazatan average annual precipitation is 503 mm. Because the monthly precipitation pattern at Mazatan follows the precipitation distribution pattern at Tombstone, Arizona, with a dominance of precipitation in July-September, and lower winter maximums during December-March, precipitation data from Walnut Gulch were

Table 2 Matape watershed subwatershed characteristics.

Subwatershed	Size (ha)	Main channel length (km)	Average land slope
1	1 607	4.1	0.10
2	1 618	8.2	0.10
3	3 647	10.8	0.10
4	5 614	5.9	0.05
5	60 899	33	0.05
6	38 674	31.6	0.05
7	67 156	37.2	0.05
8	95 192	32.5	0.03
9	30 475	39.5	0.03
10	9 035	8.8	0.03

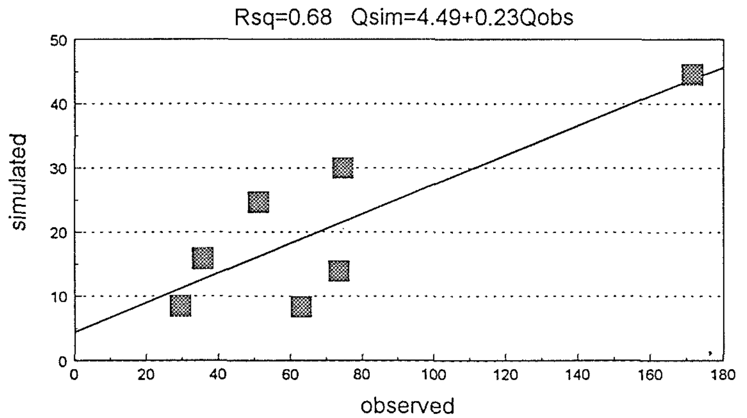

Fig. 3 Walnut Gulch Watershed observed and SWRRBWQ simulated maximum peak flow ($m^3 s^{-1}$) 1967-1977, subwatersheds 1, 3, 7, 8, 9, 10 and 15.

used as input to model the Matape watershed. The SWRRBWQ model includes a rainfall adjustment factor that is used to multiply precipitation input values by a constant for each subwatershed. By applying rainfall adjustment factors to the weighted daily precipitation data from Walnut Gulch, calculated average annual precipitation agreed well with observed precipitation for the three subwatersheds on the Matape watershed where precipitation was recorded (subwatersheds 4, 5, and 10).

Table 3 Comparison of observed and SWRRBWQ simulated peak rates and average annual runoff for the 1967-1977 period of record.

		Peak rates ($m^3 s^{-1}$):			Average annual runoff (mm):	
		Mean	Std. dev.	Max.	Mean	Std. dev.
Flume 1	observed	68.41	50.89	171.51	3.17	1.42
	simulated	7.30	8.66	44.57	3.19	3.18
Flume 3	observed	8.45	10.04	29.43	4.72	6.04
	simulated	2.15	2.49	8.48	4.55	6.84
Flume 7	observed	24.66	27.11	73.46	3.94	3.9
	simulated	3.91	3.79	13.89	4.05	7.13
Flume 8	observed	22.37	16.44	50.99	7.87	6.4
	simulated	2.21	4.08	24.67	6.7	7.76
Flume 9	observed	29.78	26.62	74.64	8.7	6.31
	simulated	3.27	4.96	29.99	8.7	7.69
Flume 10	observed	16.84	18.27	62.92	4.4	4.03
	simulated	1.46	1.90	8.39	4.26	4.24
Flume 15	observed	16.69	11.86	35.64	4.78	3.28
	simulated	4.88	4.68	15.89	4.81	6.14

Average monthly temperature, as well as maximum and minimum temperature, recorded at Mazatan were used to simulate daily temperature for each subwatershed. Monthly solar radiation values were calculated based on the latitude of the watershed. Soil parameters for the Matape watershed were estimated based on limited field data and soil texture. Because the model is sensitive to many of the soil parameters, additional data collection to accurately characterize soils is needed.

Limitations from available data on the Upper Matape watershed

Lack of observed data limited model calibration to annual water yield. Although a gauging station is located approximately 9 km upstream from the Fransisco L. Alatorre Reservoir, measured annual water yield data collected at the station is limited to eight discontinuous years of record (1960-1962, 1977, 1982-1985) with an average of 9.77 mm and standard deviation of 1.63 mm. By varying the runoff curve numbers, return flow travel time, and soil parameters over a reasonable range of values, a simulated mean annual water yield of 9.75 mm and standard deviation of 11.15 mm were calculated. The simulated water yield is based on precipitation input for the 11 years from 1967 through and including 1977 when high amounts of precipitation were recorded. If simulated water yield for 1977 is not included in the analysis, the standard deviation of simulated annual water yield is 3.08 mm and is in closer agreement with the standard deviation of observed water yield. This result indicates that caution must be observed when using short records and sparse data for calibrating the model.

DISCUSSION

The SWRRBWQ model was calibrated to estimate water yield on the Walnut Gulch and Matape watersheds. Although the large size of the Matape watershed prohibits intensive data collection, comparative watershed studies provide the opportunities to transfer data from Walnut Gulch and to use the limited physiographic data on ungauged watersheds.

Once calibrated, a model like SWRRBWQ is a useful tool for evaluating the effects of management systems on water yield. The results of this modelling effort indicate that benefits of using the model to evaluate alternative management practices will depend on the collection of additional data such as precipitation, water yield, and soil parameters. Because sediment yield is an important criteria in evaluating the impact of management systems, additional modelling must be done to accurately model peak flow and thus sediment yield.

Continuing research to calibrate and validate the SWRRBWQ model will include simulations for time periods longer than 11 years. Future modelling plans include calibration and validation on a time step shorter than one year, such as monthly. In addition, individual components of the hydrologic balance such as evapotranspiration, transmission losses, and percolation will be evaluated in detail. Efforts to collect, organize, and validate topographic, climate, soils, and hydrologic data for the Matape watershed are ongoing. After the model is successfully calibrated to estimate water and sediment yield, crop and land use data will be included to evaluate alternative management practices on rangelands in the semiarid Southwestern United States and Mexico.

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Debris flows in northeastern Victoria, Australia: occurrence and effects on the fluvial system

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Abstract A large storm in October 1993, in northeast Victoria, produced more debris flows (about 30) than have been reported before in Australia. Such debris flows are associated with storms of at least 50 year recurrence interval. The debris flows were typically between 0.5 and 1 km in length, experienced flow velocities up to 10 m s^{-1}, and were most common on resistant granite and acid-volcanic lithologies with slopes greater than about 25°. The debris flows were of little significance to modern stream processes. Thirteen debris flows in one 35 km^2 drainage basin contributed about 10% of the total sediment mobilized during the flood. Five hundred metres of channel migration and widening in the stream contributed the same volume of sediment as the largest debris flow (about 2000 m^3). However, in the longer term the debris flows are a significant process because they deliver coarse colluvium (up to 1.5 m B-axis) directly to the modern stream and flood plain, and could produce lowering of the landscape at a rate of 5 m per million years.

INTRODUCTION

Record rainfall in northeastern Victoria, Australia (Fig. 1), on 4 October 1993 produced record floods in the larger streams. For example, over three hours the Broken River rose from a minor flood to the largest flood on record (Bureau of Meteorology data). The floods caused considerable erosion, with several of the smaller tributaries suffering dramatic widening and incision. Over $4 M (Australian) of Natural Disaster Funding has been allocated to the repair of stream erosion in the Ovens and Broken River drainage basins.

An unusual feature of the storm event was the initiation of more than thirty debris flows, which are known locally as "mud flows". Several of these flows cut roads. As we describe below, these flow events have many of the classic features of debris flows (Costa, 1984, 1988), and are clearly different from other mass-wasting phenomena reported in southeastern Australia, such as the relict periglacial scree slopes and "rock rivers" of the highlands (Talent, 1965); or the earth flows (slip-circle failures) in the Otway and Strzelecki Ranges. There are few reports of debris flows in Australia.

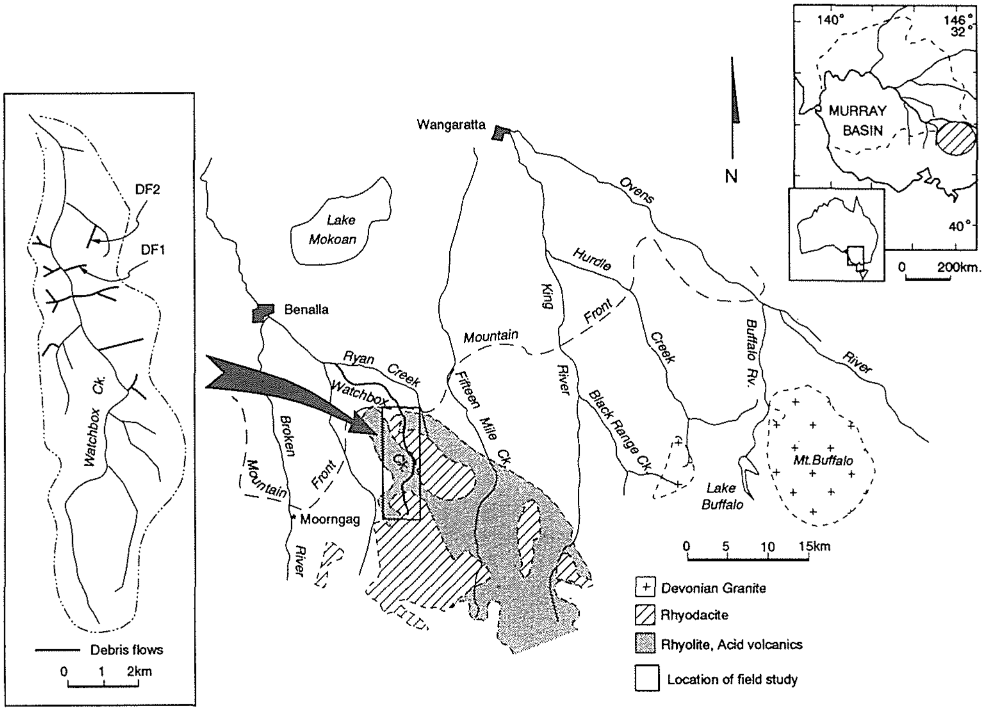

Fig. 1 Location of the study area in northeast Victoria, with detail of the Watchbox Creek basin.

Several debris flows have occurred on Mt Bellenden Kerr in wet-tropical Queensland. Wasson (1978) has described a single debris flow near Lake George in New South Wales, and another 2 km long debris flow occurred in Lilydale, Melbourne, in the 1890s (Shire of Lilydale, 1993). Thus, to our knowledge, this northeast Victorian flood event has produced an unprecedented number of debris flows in the European history of Australia.

In northeastern Victoria the debris flows are prominent effects of the floods, being visible from tens of kilometres away, and they are considered locally to have been major sources of sediment to the stream systems. This paper reports the characteristics and distribution of debris flows from the October 1993 storm event, specifically in the headwaters of the Broken River. In particular we:

(a) determine the distribution and frequency of debris flow events;
(b) describe the debris flows using two detailed case studies; and
(c) identify their importance in contributing sediment to stream systems, and their role in longer term landscape evolution.

Distribution of debris flows from the storm event

The storm event of 4 October triggered approximately 30 debris flows in the drainage basins of the Broken and Buffalo Rivers. The number of debris flows could be roughly counted from an aerial inspection after the floods. No official aerial photographs have been taken of the river headwaters since the flood. All of the debris flows occurred in forested drainage lines that have only ever been selectively logged.

Debris flows typically occurred on steep, resistant rock types with slopes above 25°. Thus, the highest density of debris flows (approximately twenty were counted) occurred on the steep and resistant acid volcanic lithologies and Middle Creeks in the drainage basins of Ryans (Fig. 1). These debris flows cut four roads, and were a major immediate impact of the storm. The highest density of debris flows (15) occurred in the Watchbox Creek basin (Fig. 1).

There were also many debris flows on Mt Buffalo (Fig. 1), a granite pluton with almost vertical cliff walls. Debris flows were restricted to the eastern and southern aspects of the mountain, particularly in Sandy, Bunyip and Boulder Creeks. Boulders from debris flows choked Sandy Creek, causing a channel avulsion. This was one of the major impacts of debris flows in the region because the new avulsed stream-channel cut through prime tobacco land.

Finally, debris flows were less common on the most common lithology in the region, Devonian greywacke and sandstone. Only 10 debris flows were counted, with two occurring in the basins of Ryans and Black Range Creeks, and smaller flows along Boggy and Fifteen Mile Creeks.

Frequency of debris flows in the region

The recurrence interval of debris flows in the region was estimated by inspecting aerial photographs from 1963, 1975, and 1993. An area of about 750 km^2, covering all of the acid volcanic lithologies (basins of Ryans and Fifteen Mile Creeks), and some of the sedimentary lithologies was selected. It was hypothesized that debris flows, even 10 years old, would be visible on the photographs, and so their recurrence interval could be estimated. The granite region of Mt Buffalo was not considered in this analysis.

No relict debris flows could be identified on the sandstone lithologies. Thus the recent debris flows on this rock-type can be considered as rare events over the last 30 years. The only relict debris flows that could be identified on the aerial photographs occurred in a small portion of the acid volcanics, with eight flows in the basin of Watchbox Creek and five in a small basin to its west. All of the debris flows could be identified on all three series of photographs. They occurred before 1963, and thus had not been covered by vegetation over 30 years.

Therefore, the last debris flows in the region occurred more than 30 years ago, and these were more spatially confined than the recent event. Engineers from the Shires of Benalla and Oxley concur that there have been no debris flows in their shires in living memory. It is also important to note that the eight debris flows in Watchbox Creek that are visible on the 1963 photographs were freshly stripped-out again in the 1993 storm, and another five new flows occurred. In short, debris flows are rare in the region affected by the 1993 flood. It probably requires a storm of at least 50 year recurrence interval to produce debris flows throughout the acid volcanic region, and to produce any debris flows in the less-steep sandstone drainage basins. Debris flow scars remain visible for at least 30 years. Further evidence for the infrequency of debris flows is provided by the stratigraphic sequences exposed where creeks have cut laterally into alluvial fans deposited below debris flow chutes. We saw no evidence in the stratigraphy of the large, sub-angular boulders carried by the recent debris flows, again suggesting that such flows are rare events.

Description of debris flows

The remainder of this paper will discuss debris flows that occurred on the acid volcanic lithology of the Watchbox Creek drainage basin (Fig. 1). Thirteen debris flows occurred in the headwaters of Watchbox Creek (35 km^2 drainage basin), a tributary of Ryans Creek, which is itself a tributary of the Broken River (Fig. 1). The closest pluviograph to Watchbox Creek is 15 km to the west at Moorngag. This station has a 12 h duration, 50 year average recurrence interval rainfall of 7.5-8 mm h^{-1} (*Australian Rainfall and Runoff*, vol. 2., map 5.8). Bureau of Meteorology records for Moorngag on the night of 4 October showed a 12 h duration of 10 mm h^{-1}, suggesting a greater than 50 year recurrence interval for the storm event. Unofficial raingauges suggest that the storm was considerably more intense than this. For example, Mr Person's 250 mm raingauge in Watchbox Creek overflowed after 5 h between 2.00 a.m. and 7.00 a.m. on 4 October. This rainfall intensity of 50 mm h^{-1} is supported by other landholders in the region.

The drainage basin of Watchbox Creek is composed of Devonian rhyodacite, rhyolite and acid volcanics. All of the debris flows occurred in fully forested drainage basins (open forest of broad and narrow-leaf peppermint (*Eucalyptus dives* and *E. radiata*)) that have only ever been selectively logged. Thus, unlike many of the debris flows described in the literature, that occur following logging (see DeRose *et al.*, 1993; and review in Gresswell *et al.*, 1979) or grazing (Lehre, 1982), these are examples of geomorphic events occurring with their natural frequency. Two debris flows were investigated in detail (described here as DF1 (grid ref. DV267383) and DF2 (DV273393)) (Fig. 1), and three others were inspected. DF1 and DF2 will be described in detail, and then compared with the other flows.

Description of debris flow 1 (DF1)

DF1 (sub-basin area of 15 ha) has left a clear, bedrock channel about 700 m long and over 5 m wide (Figs 2 and 3). The maximum measured slope in the debris flow was 33°. Sediment was deposited in a fan adjacent to Watchbox Creek. The flow had many of the classic features of a debris flow, namely: poorly sorted, coarse levee and fan deposits (both matrix and clast supported); mudlines on marginal trees; prominent super-elevation of mudlines at bends; and removal of all trees in the centre of the flow-path (Costa, 1984, 1988). The failure could be more correctly classified as a debris flow that passes downslope into a "channel confined debris torrent" (Kelsey, 1982).

Unlike other debris flows (cf. Tsukamoto *et al.*, 1982; Benda, 1990) DF1 was not initiated from a colluvial hollow, but from a straight section of the catena, with planar cross-slopes. The head of the debris flow evacuated saprolite and core-stones directly from the weathered bedrock. Hence, the heads of these debris flows represent a direct path of downslope transport of the weathered material. It was not clear whether the debris flow was initiated by failure at the top that carried the remainder of the downslope material along, or by failure at the bottom that triggered failures progressively up the slope.

The debris flow had a maximum discharge of 145 $m^3 s^{-1}$, and a maximum velocity of about 9 m s^{-1}. Flow velocity was estimated with three independent methods (Costa, 1984) (Table 1). Superelevation was measured at three bends (Fig. 2), and velocity was

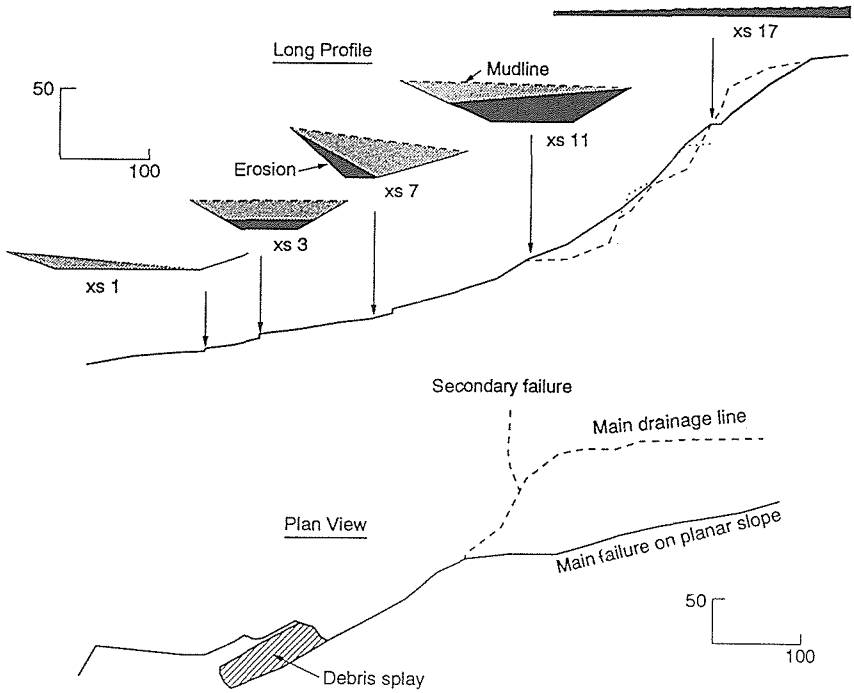

Fig. 2 Profile and plan-view of debris flow 1 (DF1). A sample of 5 of the 17 cross sections is shown.

estimated with the formula:

$$v = \sqrt{\Delta h g r_c \cos S / b} \tag{1}$$

where Δh = superelevation, $g = 9.81$, r_c = radius of curvature, S = channel slope, b = water surface width. In addition, velocity was estimated by measuring the diameter (d) of the five largest clasts ($v = 0.18d^{0.49}$), and by measuring the height of the stagnant head on the upstream and downstream side of trees

$$v = \sqrt{2gh/\alpha} \tag{2}$$

(where h = height of stagnant head, α = momentum correction factor).

There are problems using these equations, derived for normal water flows, for estimates of velocity in hyperconcentrated, non-Newtonian debris flows. Nevertheless, the results (Table 1) are very consistent at the order-of-magnitude level across the methods, indicating that the flow travelled between 4 and 9 m s^{-1}. Assuming that the erosion originated from the apex of the failure (Fig. 2), these velocity figures suggest that the front of the debris flow reached the apex of the fan about 1.5 minutes after initiation.

The instantaneous discharge carried by DF1 was probably over half of the peak discharge carried by Watchbox Creek during the flood. From the area and velocity estimated at cross section seven of DF1 (Fig. 1), discharge was estimated to be between 35 $m^3 s^{-1}$ and 95 $m^3 s^{-1}$. The lower estimate assumes that the debris flow only occupied the area evacuated by erosion, whilst the upper estimate assumes that the flow

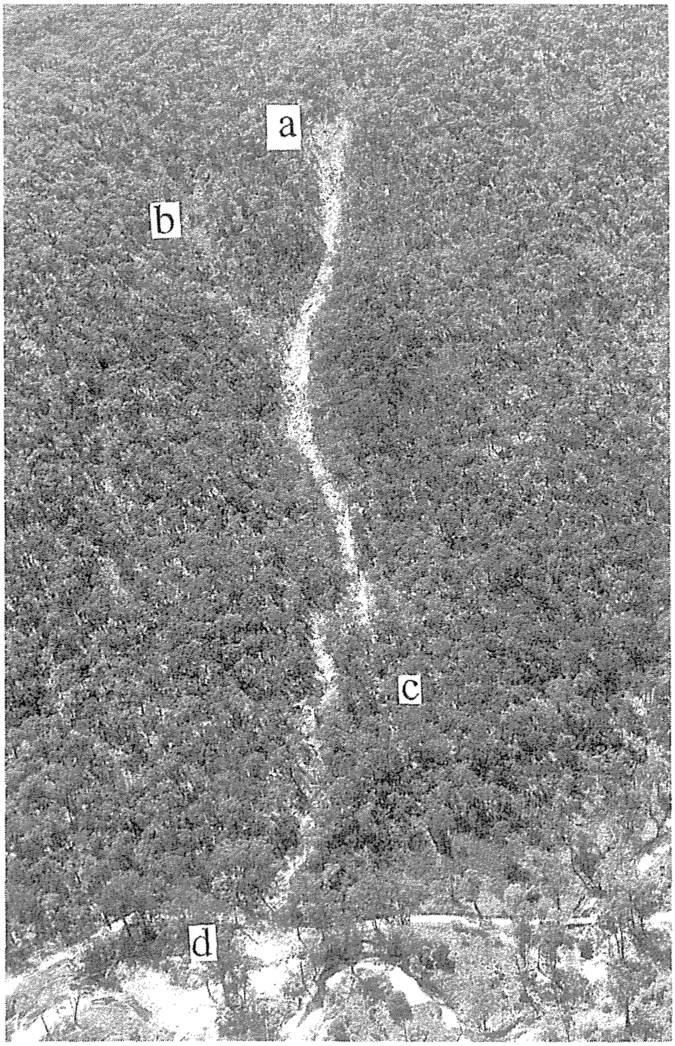

Fig. 3 Oblique aerial photograph of DF1. Note the following: the rotational slump at the head of the failure (point "a"); the main path of failure does not follow the main drainage line (point "b"); the debris splay visible through the trees below the sharp bend (point "c"); and the fan of material at the junction with Watchbox Creek (point "d").

occupied the full cross section delimited by the mudlines. For comparison, the peak discharge carried by Watchbox Creek during the flood, was estimated to be 110 $m^3 s^{-1}$ from Manning's equation. If basin area is considered to be proportional to discharge from the basin, then the discharge of Watchbox Creek can be used to estimate the water discharge that would normally be expected from the 15 ha basin of DF1 in the 1993 storm. Proportionally the basin area of DF1 should produce less than one cubic metre of water per second. Given the minimum discharge in DF1 of 35 $m^3 s^{-1}$ suggested above, it is likely that the debris flow would have contained less than 5% water, with the rest being sediment. The following approximate sediment budget suggests that less than a third of this large volume of sediment reached Watchbox Creek.

Table 1 Estimates of flow velocity in the debris flow.

Method	Position	Estimate
Superelevation	Cross section 14 (near the top) (Fig. 2)	8.9 m s^{-1}
	Cross section 7 (middle)	5.0 m s^{-1}
	Cross section 1 (apex of fan)	4.2 m s^{-1}
Five largest clasts	Bed of debris flow scar (mean diameter = 1.01 m)	5.4 m s^{-1}
	Debris splay (see below) (mean d = 1.4 m)	6.3 m s^{-1}
Stagnant head	Edge of fan (head = 2 m)	6.2 m s^{-1} (probably over-estimate)

Volume of sediment eroded and deposited

Sediment volumes were estimated from 17 cross sections surveyed across the debris flow (of which five examples are reproduced in Fig. 2). The area of material eroded at each cross section was estimated by projecting a line from the upper edge of the freshly eroded colluvium or rock, down to the edge of the old channel bed. The edge of the darker, organic-stained surface of the former channel bed could be differentiated from the freshly exposed rock surface. The proportion of silts and clays, and sand, at each cross section was estimated and compared with the grain-sizes deposited in different sediment stores in order to estimate the calibre of the material that was delivered to Watchbox Creek.

The areas of the 17 cross sections indicate that an order of magnitude more sediment was eroded from the channel margins as the debris flow passed down the drainage line, than was eroded from the alcove at the top of the debris flow (Table 2).

Sediment was deposited in four locations, in:

Table 2 Rough sediment budget for debris flows 1 and 2. Data come from field surveys as described in the text.

Zone (Defined in Fig. 2)	Erosion (m^3):		Deposition (m^3):	
	Debris Flow 1	Debris Flow 2	Debris Flow 1	Debris Flow 2
Erosion alcove at top of failure	500	750		-
"Barrel" of debris flow	7000	1500	190	500
Fan apex		1100		1500-2600
Debris splay	-		1400	-
Fan			3400	450
TOTALS	7500	3350	4990	2450-3500
Throughput to trunk stream			2510	0-900
% throughput			33%	0%-27%

(a) boulder berms within the eroded debris flow channel;
(b) coarse marginal levees;
(c) a "debris splay" (see below); and
(d) in the distal fan.

The majority of deposition (45%) was in the fan (Table 2), but a large volume (19%) was also deposited in an interesting feature which may be defined as a debris splay. Five hundred and fifty metres from the apex the debris flow met a sharp bend in the channel (Figs 2 and 3). This bend produced 2.5 m of superelevation, and as a result about 1400 m^3 of sediment was deposited over-bank on the un-failed slope beside the debris flow channel. The resultant splay was 110 m long, 20 m wide, and up to 1 m thick. It consists of a full range of particle sizes from silts and sands to 1.5 m diameter boulders (B axis).

This deposit is significant because it represents a mechanism for depositing large volumes of coarse material well above the channel. In this case the debris splay was over 6 m above the channel floor. The debris splay was deposited on a poorly sorted unit that may represent former debris splays. An initial bend in the path of a debris flow can thus produce debris splays that progressively build upward, producing a significant store of sediment in the valley.

Importantly, less than one third of the sediment eroded from DF1 reached Watchbox Creek (Table 2). The boulders deposited in Watchbox Creek from the debris flow could be differentiated from the existing bed load in the creek because they were more angular and iron stained. Although large boulders, up to 1.6 m in diameter, were deposited in the creek by the debris flow, none of the deposited particles travelled more than 100 m down the creek.

In order to compare the sediment yielded from DF1 with the sediment remobilized from the flood plain, the volume of sediment eroded by cutbank erosion in a 450 m stretch of Watchbox Creek was measured above and below the debris flow. The minimum volume of erosion was estimated from the length of tree roots exposed by the erosion. In this reach, 1800 m^3 was eroded from the flood plain by lateral erosion (58% gravel, 42% silt and clay). This was over two-thirds of the maximum 2500 m^3 of sediment delivered to the creek by DF1. Thus, about half a kilometre of bank erosion in the third-order stream mobilized as much sediment in the stream as was delivered by the largest debris flow in the valley.

This result can be extrapolated to the full length of Watchbox Creek. Of the 15 debris flows that occurred in the Watchbox Creek basin, DF1 was the only one where the deposited fan reached the trunk stream. Thus the maximum sediment yield to Watchbox Creek from debris flows would be less than twice the yield from DF1 alone, say 5000 m^3. Extrapolating the rate of flood plain erosion estimated in Watchbox Creek above, to the full 10 km length of Watchbox Creek suggests that at least 40 000 m^3 of sediment was liberated from the flood plain during the flood. This figure ignores channel incision. Therefore, it is unlikely that the debris flows introduced more than 12% of the total mobilized sediment in Watchbox Creek during the flood event, ignoring other diffuse basin sources.

Description of debris flow 2 (DF2)

DF2 occurred to the north of DF1, on the opposite side of the ridge, and flows into a small tributary of Watchbox Creek that we have named Teachers Creek. Erosion and deposition were estimated from 16 cross sections just as in DF1. DF2 is about half the length of DF1 (450 m), but shares the following features.

(a) The debris flow was initiated on a planar portion of the slope (with a slope >25°), rather than in a drainage line or colluvial hollow. Thus, DF2 removed weathered material directly from the slope. The coarsest fraction of this material came from the head of the failure (>1.5 m boulders), whilst the bulk of the sediment came from colluvium at the margins of the flow.

(b) The 3300 m^3 of sediment removed from the slope was deposited in an alluvial fan, and only a small proportion of the fine fraction was delivered to Teachers Creek (Table 2). Widening and deepening within Teachers Creek produced an order of magnitude more sediment than did erosion from DF2. Further evidence that DF2 had little impact on the stream comes from channel dimensions in Teachers Creek. A survey of 26 cross sections along the creek, upstream and downstream of the toe of the DF2 fan demonstrated that the input from DF2 had no influence on either channel size or channel width-depth ratio.

(c) DF2 delivered coarser sediment to the flood plain/fan than do other processes. Adjacent streams could transport particles of less than 0.3 m diameter, whilst DF2 delivered 1.6 m boulders.

One contrast between DF1 and DF2 is that much of the sediment eroded from DF2 came from vertical stripping of the A horizon over a large area.

DISCUSSION

The debris flows initiated by the 1993 flood in northeastern Victoria are typical of debris flows described throughout the world in terms of slopes, depth of colluvium, and flow velocities (e.g. Kelsey, 1982; Tsukamoto *et al.*, 1982; Benda, 1990). The only possible differentiating characteristic is their initiation in planar portions of the hillside, rather than in colluvial hollows.

Debris flows are rare events in Australia, with this event in northeastern Victoria being the largest debris flow event described. It should be emphasized that all of these debris flows occurred in essentially undisturbed forested basins, and probably represent a "natural" geomorphic event. Therefore, we can conclude that debris flows will only be widespread in northeastern Victoria in high intensity storms of greater than 50 year recurrence interval. They are most likely to occur in highly resistant and steep lithologies, such as granites and acid volcanics. These lithologies do not make up a large proportion of the Australian Highlands. For example, they occupy less than 10% of the Broken and Ovens River basins upstream of the mountain front.

Despite being spectacular events that leave prominent scars on the landscape, the debris flows in northeastern Victoria deliver only a small volume of sediment directly to the stream system, and they cannot be considered a management problem in terms of increased stream sediment loads. The only damage done by debris flows was to deposit boulders on roads.

Even though debris flows in this region do not deliver large volumes of sediment directly to streams, in comparison to the sediment mobilized from the flood plain storage, they are geomorphically important. This is because they transfer coarse colluvial material, derived from resistant lithologies, directly into stores of sediment that are accessible to erosion by larger streams. Thus, both DF1 and DF2 deposited coarse sediment in fans that will be eroded by lateral migration or avulsive channel change of the trunk stream. Other erosion processes move material more slowly. For example, soil creep will move coarse sediment directly onto the flood plain, whilst gradual erosion by tributary streams can move some coarse material onto fans. The largest clast that was observed to be delivered to the trunk stream by the floods, without a debris flow, was 0.3 m in diameter. By contrast, DF1 delivered 1.5 m boulders to Watchbox Creek. Of course, it is not clear whether the gradual processes of creep and stream erosion move a larger total volume of material than do the catastrophic, but rare, debris flows.

The debris splay reported here from DF1 is an interesting geomorphic feature. In this case, superelevation of the debris flow has been a process that can deposit very coarse material over 5 m above the channel floor. As the deposit grows deeper with successive debris flows it will become more effective at diverting the flow, thus producing a feedback mechanism that could lead to the deposition of several metres of sediment within a debris flow chute.

Finally, it is interesting to consider the possible role of debris flows in sculpting headwater slopes. All of the debris flows inspected in the northeast stripped the regolith to bedrock, and large areas of fresh bedrock were exposed where rocks in transit had struck the surface. In particular, the debris flow tended to smooth the floor of the channel by abrading any irregularities or protrusions. Although this form of erosion appears minor during one debris flow event, over millions of years it is significant. For example, in the upper end of DF1, about 5% of the bedrock surface was chipped to at least 10 mm depth. If it is assumed that debris flows occur every 100 years, then by extrapolation, approximately 10 mm would be removed from the floor of the flow every 2000 years. Assuming the frequency of debris flows remained the same, then over the 2 million years of the Pleistocene debris flows could have eroded the drainage line a distance of 10 m. This would translate to a lowering of the basin divide by 5 m per million years, which is consistent with long-term denudation rates throughout south eastern Australia (Bishop, 1985; Gale, 1992).

A weakness in this extrapolation is the possibility that debris flows become less frequent as the drainage line is eroded. Debris flows in the northeast did not occur in drainage lines that were deeply incised. That is, there were no debris flows where the side slopes into the drainage line were very steep. In these drainage lines the colluvial material is removed before it builds up in the stream, or on its margins. This could be in part related to the size of the basin, but it could also be related to the more efficient delivery of colluvium to the drainage line by the steeper side-slopes. This would imply that debris flows become less frequent as the drainage line is incised. This proposition has implications for the relationship between the evolution of stream channels and the evolution of the catena into the stream. Unfortunately the sample size of debris flows in this region is probably insufficient to test the proposition quantitatively.

CONCLUSIONS

A large storm in October 1993, in northeast Victoria, produced more debris flows (about 30) than have been reported before in Australia. Such debris flows are associated with storms of at least 50 year recurrence interval. The debris flows were typically between 0.5 and 1 km in length, experienced flow velocities up to 10 m s^{-1}, and were most common on resistant granite and acid-volcanic lithologies with slopes greater than about 25°. These characteristics are typical of debris flows that are common elsewhere. The only possible differentiating characteristic of the debris flows investigated on the acid-volcanic lithologies is their initiation in planar portions of the hillside, rather than in colluvial hollows.

The debris flows were of little significance to modern stream processes. Thirteen debris flows in one 35 km^2 drainage basin contributed approximately 10% of the total sediment mobilized during the flood. Five hundred metres of channel migration and widening in the stream contributed the same volume of sediment as the largest debris flow (approximately 2000 m^3). However, in the longer term the debris flows are a significant process because they deliver coarse colluvium (up to 1.5 m B-axis) directly from the hillside to the modern stream and flood plain, making it available for later transport. In addition, erosion of bedrock by the high velocity debris flows could produce lowering of the landscape at a rate of 5 m per million years. It is probable, however, that the frequency of debris flows declines as drainage lines incise.

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5 Human Impacts

Variability in Stream Erosion and Sediment Transport (Proceedings of the Canberra Symposium, December 1994). IAHS Publ. no. 224, 1994.

Erosion and sediment yields in the Kakadu region of northern Australia

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Abstract The undisturbed landscape of the Kakadu region of northern Australia is highly stable despite it's geographic position in the wet/dry tropics; a world zone which exhibits the highest rates of water erosion and sediment yield. Measured sediment yields in the Australian region are low by world standards and are comparable to the low rates typical of the Australian continent in general. In the Kakadu region, erosive rainfall is countered by erosion resistant landforms; the rocky plateau surfaces and the gravel armoured lowlands. On the lowland slopes, the litter cover formed during the prolonged dry season also plays a role in limiting soil erosion but vegetation cover in general is not as important as the surface lag gravel in limiting erosion. Once the lag is disturbed, erosion rates increase by 2 orders of magnitude and rapid gully erosion occurs even on very gentle slopes. The lag can be encouraged to reform on disturbed lowland slopes through reformation of the former contours and use of local, gravely topsoil. Rehabilitation programmes based on this approach are likely to be more efficient and persistent than the Australian standard methods involving initial establishment of a dense exotic vegetation cover.

INTRODUCTION

The Kakadu region of Australia is part of an extremely old landscape in an ancient continent. In this landscape, weathering and erosion over many millennia have produced intricate rocky escarpments and vast lowland plains. It is an outwardly stable landscape in which landforming processes, although many and varied, are generally slow and insidious.

If the Kakadu region is unspectacular in any respect it is in relation to this geomorphological activity. In the undisturbed environment, processes such as soil erosion and sediment transport are active but very weak. The most spectacular landforming events are perhaps the isolated rock falls from scarp surfaces.

The stability of the Kakadu landscape is unusual considering its position within the wet/dry tropics. In this zone, the annual coincidence of drought ravaged vegetation cover following the prolonged dry season, with the onset of intense wet season rains, produces amongst the world's highest natural rates of soil erosion.

This paper examines why the landscape of Kakadu is so stable and why it is nonetheless highly susceptible to development pressures. It also proposes that conventional soil conservation techniques will not necessarily assist in stabilizing land used intensively in this region. It is argued that the development of land management practices which *are* suited to the Kakadu region depends on an understanding of how these practices interact with natural land forming processes.

This paper draws from a large body of soil erosion and sediment transport data collected between 1979 and 1987 in both disturbed and relatively pristine drainage basins. These data relate primarily to the impacts of land clearing associated with mining, but also generally apply to any intensive land use involving clearance of vegetation on the lowlands.

THE KAKADU ENVIRONMENT

The Kakadu region encompasses an area of over 20 000 km^2 in the northwest of Australia's Northern Territory. The region includes the basins of the East, South and West Alligator Rivers. The rivers rise in the rocky Arnhem Land Plateau in the east and flow across gently sloping, wooded lowlands and flood plains to the Arafura sea. The region includes Kakadu National Park, Australia's first proclaimed World Heritage area. It also contains abundant mineral deposits, notably gold and uranium, and a uranium mine (Ranger Uranium Mine) currently operates in the drainage basin of the East Alligator River. This mine is serviced by a settlement of around 1500 people at Jabiru.

Landforms

The major landforms of the Kakadu region are:

(a) The sandstone plateau, scarps and rocky outcrops of the geological formation known as the Kombolgie Formation. Kombolgie sandstone, composed of quartz sandstone with minor conglomerate and interbedded volcanics (Needham *et al.*, 1980), is unconformable on the Lower Proterozoic metasediments and gneissic equivalents of the Pine Creek Geosyncline (Stuart-Smith *et al.*, 1980). The Kombolgie Formation covers much of the area in the south-east of the region, forming the Arnhem Land Plateau, but has been substantially removed through dissection and scarp collapse in the north and west (Galloway, 1976). The Plateau is inclined gently eastwards but forms an abrupt scarp along the western margin, rising to 300 m above the lowlands.

(b) The gently sloping, wooded lowlands. The lowlands comprise a level to undulating, weathered terrain, supporting open forest and woodland and formed by erosion and duricrusting of the Lower Proterozoic metasediments during the Late Tertiary and early Pleistocene (Williams, 1969). The Lower Proterozoic rocks are irregularly but extensively weathered (Needham, 1972) and are capped by a ferricrete horizon which forms the base of the contemporary soil (Milnes *et al.*, 1986). Dominant soil types are single grained to massive, sandy to gravelly red and yellow earths, earthy sands and siliceous sands. A striking feature of the lowland surface is the presence of a dense gravel lag, comprising vein quartz and ferruginous gravel, forming up to 100% cover on the surface of the dominant red and yellow earths.

(c) The seasonally inundated river systems and flood plains. In the upper reaches and over the lowlands, the river systems are typically anastomosing, sandy channels fringed by woodland vegetation. These channels cease to flow during the dry season and contract to a series of semi-permanent waterholes. In the lower reaches, the wet season floods flow over extensive freshwater flood plains into large, muddy, meandering tidal channels cut into estuarine mangrove swamps. The present flood plains were formed through deposition of about 6 m of estuarine sediments during the last marine transgression (around 10 000 years ago). These are overlain by a more recent layer of fluvial muds (Woodroffe *et al.*, 1985).

Climate

The climate of the region will be dealt with in this paper in-so-far as it influences contemporary erosion and sediment transport. The two most important climatic features in this regard are the prolonged annual drought, broken by an intense wet season, and the high intensity of rainfall events. The annual drought results in a dramatic reduction in vegetation cover, leaving the soil surface exposed to wet season rains, while the high intensity of many of these rains further increases the potential erosion hazard.

The climate is characterized by a hot wet season (December-March) and a hot dry season (May-October). April and November are transitional. Temperatures are high throughout the year with mean monthly maxima of 37°C in November and 31.5°C in July (McAlpine, 1976). Mean annual rainfall ranges from over 1500 mm in the north of the region to 1300 mm in the south, and close to 90% falls in the wet season. Rainfall intensities of 100 mm h^{-1} for 10 min periods may be expected annually and cyclones recur on average once or twice per year (Lee & Neal, 1984). Extreme cyclonic rains may account for up to one third of the annual rainfall total in the space of 24 h (Jackson, 1977). At least 25 such events have been recorded in the north-west of Australia since 1915 (McGill, 1983).

The erosivity of rainfall (based on the EI_{30} index derived by Wischmeier *et al.*, 1958) in the region is high compared with the rest of the Australian continent, exceeded only on Cape York Peninsula in tropical Queensland (McFarlane & Clinnick, 1984). Based on these rainfall characteristics, the Kakadu region may be expected to experience higher than average rates of soil erosion by water.

SOIL EROSION AND SEDIMENT TRANSPORT

A study of erosion and sediment transport was carried out in relatively undisturbed drainage basins in the region between 1981 and 1987. The study involved monitoring of hillslope erosion using bounded runoff plots and erosion pins, and measurement of the sediment yields of four basins ranging in area from 5 to 16 km^2. The basins are otherwise similar in morphology, with major land systems, soils and landforms represented (Fig. 1).

Sediment yields

Monitoring of suspended sediment and solutes in the four drainage basins was carried out using automatic flow recorders and flow proportional samplers (ISCOtn). A total of 3698 samples was collected during the study period. This period included a range of rainfall years with one above average year (1984-1985 wet season), two near average

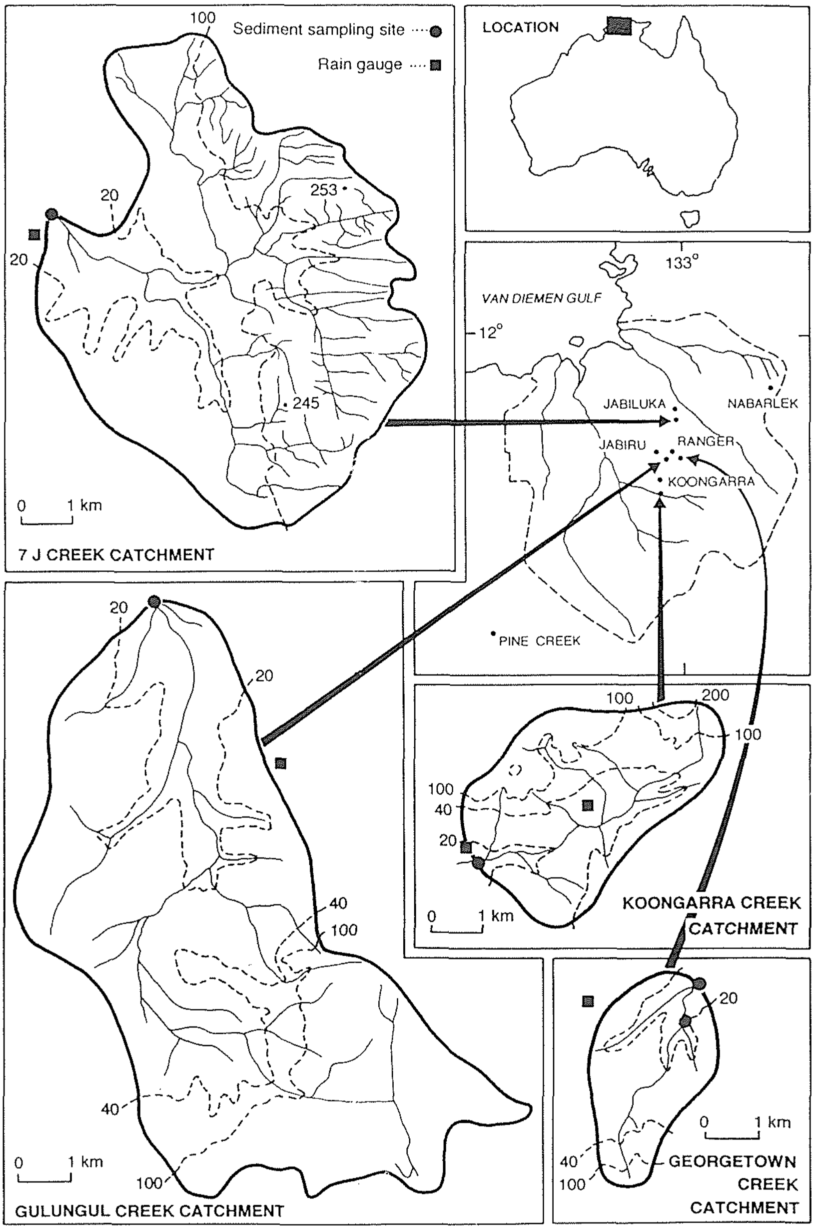

Fig. 1 Map of research basins in the Kakadu region.

years (1981-1983 wet seasons) and two below average years (1985-1987 wet seasons). The high frequency of sediment sampling permitted calculation of actual event yields, integrating sampling intervals of generally less than 1 h and commonly less than 5 min.

The annual suspended sediment and solute yields are shown in Table 1. Mean annual yields of suspended sediment (silt, clay and minor fine sand) and solutes (corrected for atmospheric inputs) for the four basins averaged over nine data years using a specific gravity of 2.5 g cm^{-3}, convert to a mean denudation rate of 12 mm 1000 years^{-1} (SD

Table 1 Sediment yields and rates of erosion and denudation for four drainage basins in the Alligator Rivers region.

Catchment	Area (km^2)	Rainfall (mm)	Discharge (1000 m^3)	Suspended yield (t)	Solute yield (t)	Annual erosion (t km^{-2})	Denudation (mm 100 $years^{-1}$)
Koongarra Creek							
1981-1982	15.4	1452	13 173	489	184	44	17
1982-1983	15.4	1206	8 216	439	126	37	14
7J Creek							
1981-1982	53.5	1451	14 413	505	274	15	6
Gulungal Creek							
1984-1985	61.9	1781	31 109	3607	504	66	26
1985-1986	61.9	967	14 227	697	276	16	6
1989+1987	61.9	1120	19 543	2163	319	40	16
Georgetown Creek 1							
1984-1085	7.8	1781	4 000	250	109	46	18
Georgetown Creek 2							
1985-1986	4.8	967	245	13	5	4	2
1986-1987	4.8	1120	864	47	20	14	6
Mean						31	12

7.8). Volumetric estimates of the bed load component of total sediment yield suggest a long term delivery rate of bed material equivalent to around 4 mm 1000 $years^{-1}$, or around 25% of the total load. The corrected solute yields account for 10 to 25% of the total load, and suspended sediment accounts for around 50 to 65%. Total mean denudation derived from the sum of suspended, solute and bed loads is estimated to be in the order of 16 mm 1000 $years^{-1}$.

Hillslope erosion

Erosion pin experiments in the lowlands over the period 1979 to 1983 provide insight to erosion processes operating on the slopes. A total of 1700 erosion pins were monitored at sites representing the major land systems of the lowlands. At each site, 100 pins (galvanized roofing nails, 10 cm in length) were emplaced flush with the surface at equal intervals along surveyed transect lines (10 m or 20 m) on the contour (a total of 27 transects of 100 pins each). The loss or gain was measured from the head of each pin at the end of the wet season. The values for the 100 pins at each site were averaged to give the mean value of annual surface lowering for the site. The pins were measured once only because of the destructive effect of measuring deposition. The records at each site are therefore for 1 year.

The variation in soil movement across the slope is illustrated for two transect sites (one burnt in the previous dry season) in Fig. 2. This illustration indicates that both

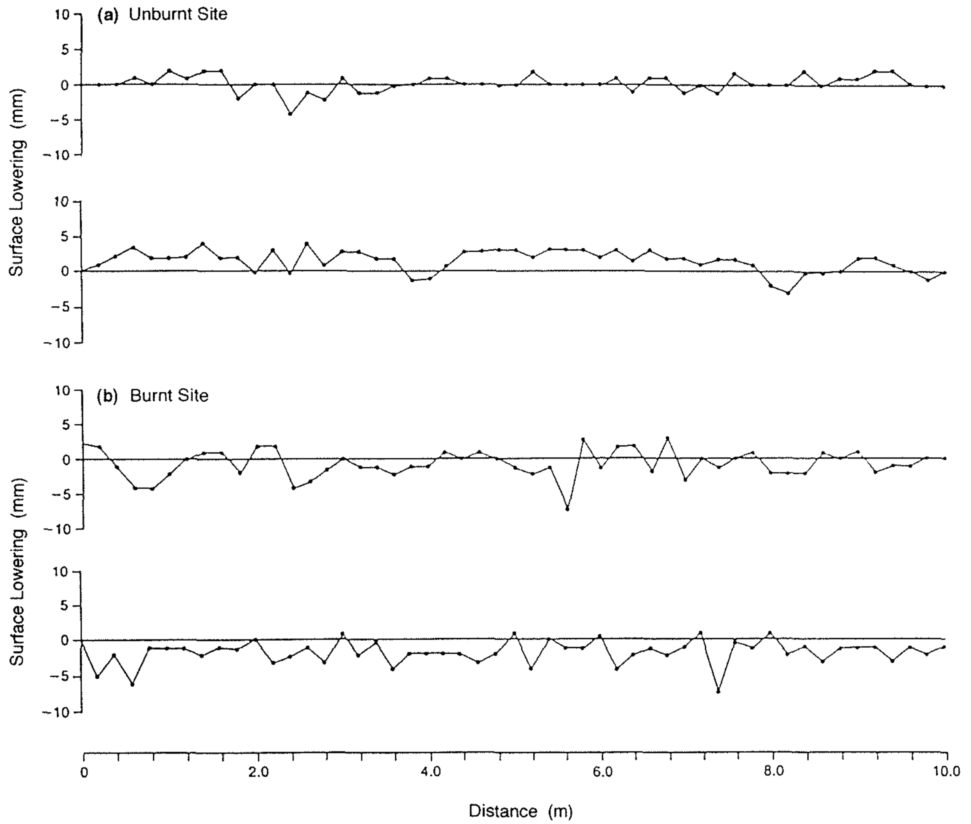

Fig. 2 Surface movement at two erosion pin transects showing net deposition (+0.94 mm) at site (a) and net erosion (−1.24 mm) at site (b). Site (b) was burnt in the previous dry season.

erosion and deposition are occurring, with net accumulation at transect (a) and net erosion transect (b) over a one year period (1980-1981). Net deposition was measured at 24 of the 27 transect sites in the lowlands (Fig. 3).

Although the erosion pin data do not indicate the periods of sediment storage on the lowland slopes, they do suggest that the path of material eroded from the Kombolgie sandstone involves a very long period of slow slope wash across the lowland slopes. Even apparently significant surface movement measured at certain pin sites may reveal nothing more than minor displacement of single gravel particles (the steep deposition peaks depicted in Fig. 2, because they are isolated, probably reflect movement of single gravel particles, which are up to 20 mm diameter).

Global perspective

The wet /dry tropics has among the world's highest recorded sediment yields and rates of water erosion (Saunders & Young, 1983). In the arid zone, erosion and sediment yields are low because of the infrequency of rainfall and runoff. They are low in the humid tropics because of the dense forest cover. In the seasonally wet tropics, however,

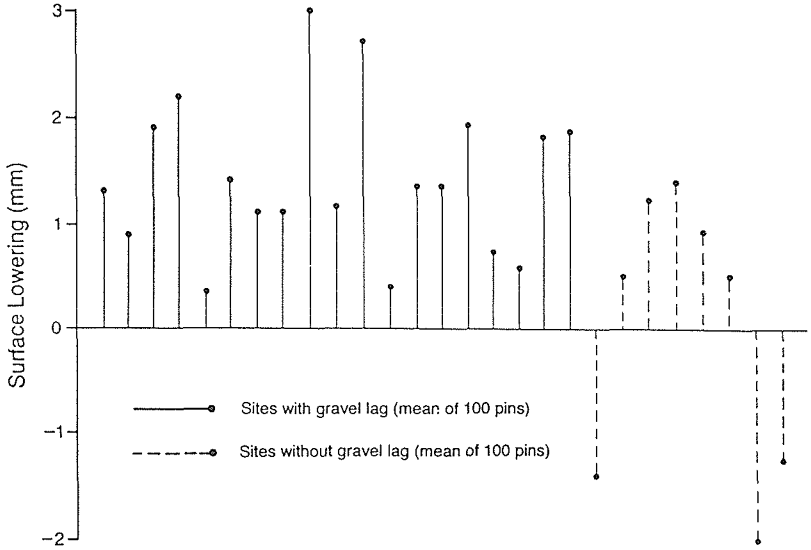

Fig. 3 Net surface movement measured over one year at 27 erosion pin transect sites in the natural lowlands of Kakadu.

high rates of erosion and sediment yield are expected because of the coincidence of sparse vegetation and intense rainfall at the beginning of the wet season. Based on rainfall intensity alone, the Kakadu region would be expected to experience high rates of denudation relative to the rest of the Australian continent (McFarlane & Clinnick, 1984).

The denudation rates estimated from sediment yields in the Kakadu region are an order of magnitude lower than the typical range reported from the wet/dry tropics (100 to 500 mm 1000 years^{-1}; Saunders & Young, 1983) and are comparable to the low rates typical of the Australian continent in general (Olive & Walker, 1982), and of forested, tectonically stable areas of the humid tropics in Australia and Malaysia (Douglas, 1976).

Factors controlling erosion and sediment supply

Although live vegetation cover is sparse, significant litter cover is present on the surface at the beginning of the wet season. Pre-wet season litter cover, even in burnt areas (resulting primarily from leaf and debris fall) is typically greater than 30% and cover of up to 90% has been recorded (mean at 27 erosion pin transect sites was 50%; Duggan, 1989).

The substantial litter cover would act to deflect rain energy and slow down runoff, thereby protecting the surface from erosion. A further influence of litter is the formation of litter dams, particularly on burnt slopes. The litter dams reduce the rate of runoff, and create a stepped microtopography conducive to deposition and storage of sediment. A weak but statistically significant inverse relationship between pre-wet season litter cover and surface lowering ($r = -0.57$; SEE 0.36; significant at the 0.01 level) was found in the erosion pin study reported above (Duggan, 1989).

The litter cover is not, however, the major factor in the stability of the lowland surface. The effect of litter is largely overshadowed by the mechanical resistance to rain splash and surface wash afforded by the surface gravel lag. The most compelling evidence for this is found in what happens when the gravel lag is disturbed.

The erosion pin data show that an increase in the rate of surface lowering may be expected following removal of vegetation cover through burning. Disturbance of the surface lag, however, produces dramatic increases in rates of erosion and rapid gully formation, even on slopes of less than 2°.

DEVELOPMENT IMPACTS

A study of the impact of land clearing at the Ranger and Nabarlek (Fig. 1) uranium mines on erosion and sediment transport was carried out between 1979 and 1987. This study examined hillslope erosion at lowland sites cleared for infrastructure, housing and borrow pits.

The mean rate of surface lowering calculated from monitoring of 2400 erosion pins on disturbed lowland slopes near the Ranger and Nabarlek mines was found to be 1.25 mm year^{-1} (SE 0.32). This is 2 orders of magnitude greater than levels recorded on undisturbed lowland slopes (Fig. 4). Soil loss on steep embankment slopes (17°) at the mine sites ranged from 2.0 to 7.0 mm year^{-1}, with the highest value being from an bare earth embankment, and the lowest value from an embankment stabilized with a rock mulch. Values for vegetated embankments fell between these extremes.

The study also involved monitoring of the sediment yields of two mine site (Ranger and Nabarlek mine sites) and one urban drainage basin (Jabiru Township). The three basins, ranging in size from 15 to 22 ha, were monitored for three to four years following clearing and construction of buildings and infrastructure. Annual suspended sediment yield was calculated from flow proportional sampling of runoff at the outlets of the basins. A total of 2049 samples was collected from 316 discharge events in the three basins between 1979 and 1983.

Annual suspended sediment yields of 2000 to 3000 t km^{-1} are recorded for the three disturbed basins during the construction phase (Fig. 5, year 1). These figures compare with yields of less than 70 t km^{-1} recorded for the four undisturbed basins reported above.

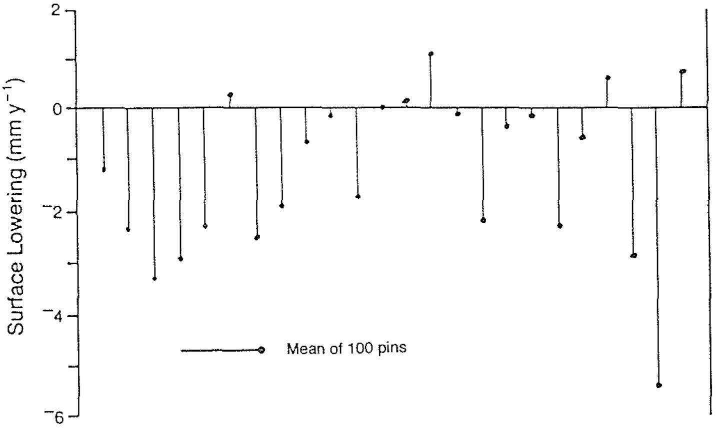

Fig. 4 Net surface movement measured over one year at 24 erosion pin transect sites on disturbed lowland slopes.

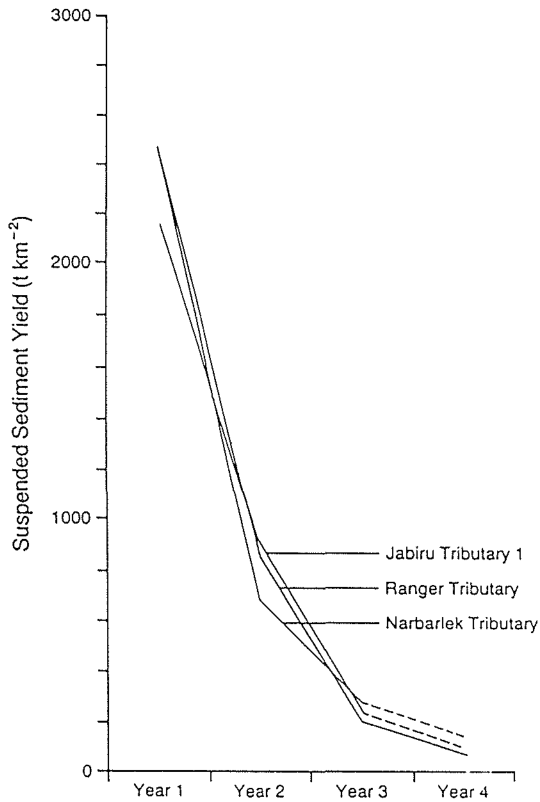

Fig. 5 Annual sediment yields of disturbed lowland drainage basins in the Kakadu region.

Not only was the impact of disturbance evident in the magnitude of sediment yields (which could be influenced by differences in basin area) but also in the nature of the sediment transported. Sediment yields in the disturbed basins comprised large proportions of silt and clay, resulting in highly turbid runoff, in marked contrast to the clear runoff from undisturbed slopes in the region.

Annual suspended sediment yields in the disturbed basins, however, fell dramatically to natural levels within four years (Fig. 5). The fall in sediment yield reflects the progressive establishment and maintenance of (largely exotic) vegetation on the disturbed slopes of the basins. This traditional form of stabilization and erosion control apparently works in the short term, but observations of erosion processes at rehabilitated sites suggest that it is not sustainable in the Kakadu region.

LESSONS FOR REHABILITATION OF DISTURBED SLOPES

Over 60% of total annual erosion on disturbed, bare surfaces occurs in the initial two months of the wet season (Duggan, 1989). Therefore, unless irrigation is used to promote establishment of the exotic ground cover before the onset of the wet season, little is achieved for significant cost in the first year of revegetation.

Much of the fertilizer applied to the rehabilitated areas is also lost during the initial wet season rains. Over 25% of nitrogen and about 90% of potassium applied at a rehabilitated site treated with standard techniques near the Jabiru Township was lost in the first wet season (Duggan, 1989). These nutrients would be washed to nearby streams with the potential hazard of nutrient build-up and associated impacts on the local ecology.

Most importantly, with the combination of dry season drought and fire, the exotic grasses do not persist for longer than about three years unless constantly maintained, and a bare surface is exposed once again to wet season rains as the exotics die out. In effect, the exotic cover established at considerable expense, only serves to delay a high rate of erosion.

The means by which long term stabilization can be achieved in the absence of intensive maintenance of exotic vegetation cover are worthy of investigation. The studies of erosion on natural and disturbed slopes reported above indicate that the primary factor limiting erosion in the lowlands is the coarse surface lag. A crude but demonstrative study of the formation of the surface lag carried out at a rehabilitated site near Jabiru Township showed that the lag will re-establish within one wet season if the surface is reshaped to pre-existing contours, and spread with very gravelly soil (Duggan, 1989). The study involved monitoring of the sediment yield of the rehabilitated site. In addition, rain splash exclusion screens (1 m^2) set 1 m above the ground surface were used to examine lag formation. The lag rapidly formed around the screens but did not form beneath them.

Neither does the lag form under a dense cover of exotic vegetation, because, like the exclusion screens, the vegetation cover absorbs rain splash energy (the main agent in formation of the lag) and prevents selective erosion of fines. While an alternative method of rehabilitation, encouraging rapid formation of a surface gravel lag, promotes a high initial rate of erosion as the lag is forming, the surface is stabilized within one wet season. The resulting stable surface is much more likely to persist in the long term than is an exotic ground cover, and is achieved at considerably less cost.

The same situation applies to the steeper embankment slopes created on mine sites. These are more likely to be stabilized in the long term through re-establishment of the configuration of natural steep slopes in the region. These are armoured by coarse rock. While vegetation cover also plays a role, it is far less significant than the rock armour. These principles need to be understood and applied to rehabilitated sites in the region.

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Variability in Stream Erosion and Sediment Transport (Proceedings of the Canberra Symposium, December 1994). IAHS Publ. no. 224, 1994.

Analysis of sediment transport developments in relation to human impacts

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Abstract The objective of this paper is to analyse the long-term development in river degradation processes in a section of an Alpine Austrian river. Due to human impacts, such as river training works, gravel mining in the main river, torrent control structures in tributaries, both the bed load transport capacity and consequently the river morphology have changed. The collected data refer to longitudinal low water level measurements dating back to 1886, to detailed river bed monitoring of cross sections, grain size distribution of sediments obtained from several tons of samples and to gravel extraction data. Unfortunately, no sediment transport measurements are available. It can be concluded that the rate of degradation is substantially caused by gravel mining and by river channelization. The effect of sediment trapping in tributary basins can be seen in the longitudinally varying rate of degradation. This analysis should provide a basis for the redesign of the cross sections to reduce the degradation process in the long term.

INTRODUCTION

Rivers are very effective transport agents, which, through associated erosional and depositional activity, are responsible for changes in the morphology of river systems (Hey, 1987). However, human impacts have serious short and long-term effects on river morphology and sediment transport.

Despite the increasing economic importance of erosion, sediment transport and sedimentation, there are still very few data from long-term detailed monitoring programmes; this situation is unlikely to change in the near future given the current economic climate (Olive & Rieger, 1992). Changes in sediment transport have often been caused by engineering measures undertaken during the last century. These have lead to degradation problems in many Alpine rivers (Hunziger, 1991; Jäggi, 1992; Nachtnebel & Habersack, 1993).

River training works resulted in the long-term in an increased sediment transport capacity, while sediment trapping in the tributary basins has reduced the sediment input into the main river. Consequently, the bed load regime was substantially modified and degradation processes became dominating (in the long-term). A similar development was observed along the River Drau, located in the southern part of Austria. To maintain river protection works against scouring, increasing efforts had to be undertaken in the last decade.

The aim of this paper is to analyse the development of sediment transport in relation to human impacts, regarding spatial and temporal variabilities within a long river reach.

The analysis of the sediment budgets should assist in the development of economically justified and ecologically sound river management strategies.

STUDY REACH

The study reach is part of the River Drau (Fig. 1), which enters Austria in eastern Tyrol and flows to Slowenia. There are two dominant geological units in the catchment which are separated by the Drau: the Central Alps in the north (granite, gneiss) and the lower limestone Alps in the south.

Concerning sediment transport, and especially bed load supply, the Central Alps in the north (including the highest mountain in Austria and some glaciered areas) supply the most resistant sediment. Thus, they are quite important for the sediment transport and armouring processes in the study reach of the Drau. In comparison to the other tributaries, the River Isel, draining a large area in the central Alpine region, has the greatest influence on the sediment regime.

Hydrological data for the Drau in the study reach are given in Table 1. The distribution of the seasonal discharge is dominated by the contribution from glaciers, with minimum in winter and a maximum in June/July. Discharges above 300 $m^3\ s^{-1}$ cause flooding, covering the whole valley, which is densely populated in several reaches.

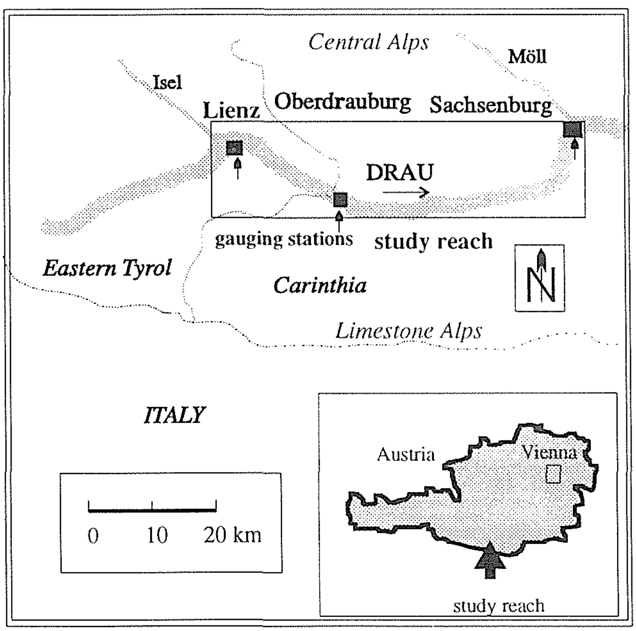

Fig. 1 Study reach at the River Drau in Austria.

Human impacts

Historical maps show that prior to 1890 the Austrian part of the River Drau was characterized by a large sediment supply from the Alpine sources. This resulted in a

Table 1 Hydrological data for three selected sections of the Drau.

Gauging station	Lienz	Oberdrauburg	Sachsenburg
Location (km)	0	15	57
Drainage basin (km^2)	1876.2	2112	2561.4
Slope ($m\ m^{-1}$)	0.0025	0.002	0.0015
Channel width (m)	30	30-40	40-50
Average flow ($m^3\ s^{-1}$)	55	65	76
Q_{30}, HQ_{100} ($m^3\ s^{-1}$)	632; 760	706; 854	840; 1029

braided, aggrading channel system. Hence, one hundred years ago engineers were required to find solutions that would reduce aggradation and, particularly, the risk of flooding. At that time the reduction of channel width and bank protection measures were regarded as appropriate solutions to stop aggradation.

From 1890 to 1930 the river was channelized to a constant, uniform channel width of 30-50 m, resulting in a 34% reduction of the previous braided area. Because bed aggradation ceased and flooding was reduced, the engineering methods seemed to be successful. After 1930, and especially after 1965/1966, when high floods occurred, further regulations, bank protection measures and torrent control structures in 19 small sediment transporting tributaries were erected. These measures, the establishment of water power plants and gravel mining, affected sediment transport and river morphology. Adverse ecological developments, such as lack of dynamic gravel bars, decreasing ground water level and reduction of "Aue" areas, as well as interruptions of the river continuum have to be observed.

DATA BASE FOR SEDIMENTS AND MORPHOLOGY

The data base includes the following elements:

(a) maps dating back to 1820;
(b) annual longitudinal low water level measurements beginning in 1890;
(c) annually repeated cross sectional measurements since 1991;
(d) data from water authorities of the Drau and its tributaries;
(e) data concerning gravel mining; and
(f) grain size distribution of subsurface material.

Maps from 1824 and 1880 depict the river reach before regulation measures were installed. Annual low water measurements began in the late 1890s and annual longitudinal profiles are available for the Carinthian part of the study reach since 1931, with exception of World War II and 1965-1967. Low water (November until February) measurements at a 200 m spacing show the temporal development of the river bed as a consequence of sediment transport.

These results can be compared to repeated annual measurements of 14 cross sections, undertaken since 1990, and regression analysis of low water level developments at the gauging stations and river bed developments at bridge piers. Gravel

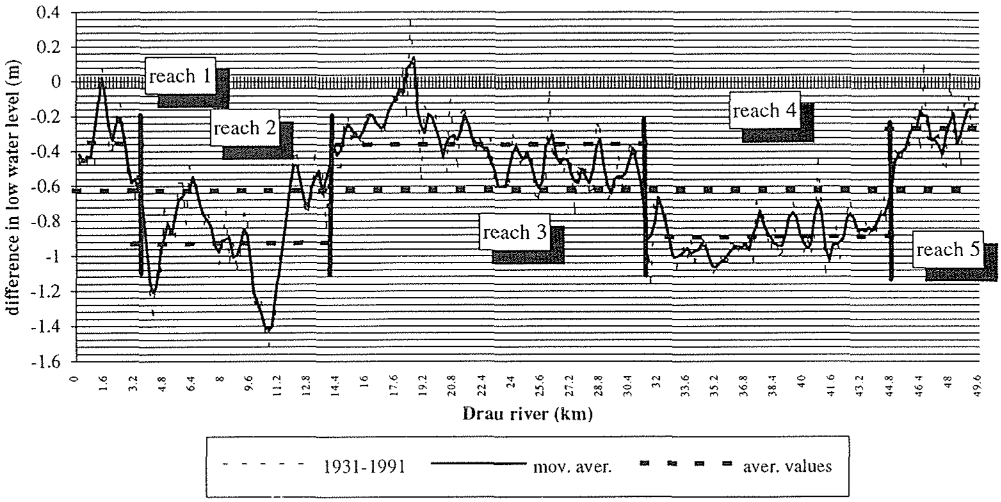

Fig. 2 Spatial variability of bed level developments at the Drau.

mining data and results of sediment analysis are compared with low water measurements to identify the different influences.

ANALYSIS OF SEDIMENT TRANSPORT

Spatial variability

Analysis of annual low water level measurements for the period 1931-1991 shows that the river bed degraded at a mean rate of 1.0 cm year^{-1}. Maximum degradation was 1.53 m (Fig. 2). Had a degradation rate of 1.0 cm year^{-1} occurred throughout the whole reach, the suggestions for further engineering methods would be not difficult. Volumetric comparisons between gravel mining and variations of the bed level showed that 60% of the changes were caused by gravel mining. In principle it can be concluded that gravel mining should be prohibited. At a few locations a surplus in sediment deposits after major floods would temporarily be available for gravel mining.

An obvious division of the Drau River in the Carinthian section into five separate, significantly different river reaches can be suggested in terms of the -0.6 m average degradation rate of the whole reach (Fig. 2). Reaches 1, 3, 5 have lower mean degradation rates (−0.32 m, −0.35 m and −0.28 m respectively), whereas reaches 2 and 4 are characterized by higher mean values (−0.95 m and −0.87 m).

Comparison of cross section measurements for the years 1990 and 1991 demonstrates that 66.9% of the whole river reach degraded, and 33.1% aggraded. This could be interpreted as reduction of the degradation tendency, evident from the low water level measurements, where 96% degraded, only 3.6% aggraded and 0.4% remained in equilibrium.

Regression analysis shows that low water level changes at the three gauging stations correspond with the low water level measurements at 200 m increments, as well as with changes of bed level observed near bridge piers. Hence, the whole analysis is highly accurate.

Besides these measurements, mathematical calculations and the application of simulation models allow the analysis of the present situation, and the prediction of developments of the river bed. For the River Drau, the bed load transport rate was calculated with the Meyer-Peter & Müller (1949) formula. In eastern Tyrol there exist levees on both sides of the Drau, which allow a flood control for HQ_{100}, whereas in Carinthia already floods of HQ_1 to HQ_5 inundate the whole valley. These different engineering methods affect the sediment transport rates (Fig. 3). In Carinthia, the transport rate corresponding to a discharge below 330 $m^3 s^{-1}$ is higher than that in eastern Tyrol. The transport rate is then rapidly stabilized at 600 $m^3 s^{-1}$ because of the large flood plain areas in Carinthia. For the transport rate in eastern Tyrol this point is reached at a discharge of 900 $m^3 s^{-1}$, when levees are already flooded.

Grain size analysis of the subsurface (volumetric samples) has demonstrated that the mean diameter in reaches with large degradation is less than that of equilibrated or slightly degraded reaches. In places this is due to small tributaries in some reaches transporting coarser material to the Drau, creating an armouring layer. In reach 2 the mean diameter of the subsurface material varied between 22 mm and 26 mm, in reach 3 between 32 mm and 33 mm, in reach 4 *c.* 20 mm and in reach 5 *c.* 33 mm. The application of the formulae of Gessler (1965) and Günter (1971) demonstrate that in the whole reach of the River Drau there is a strong tendency for armouring.

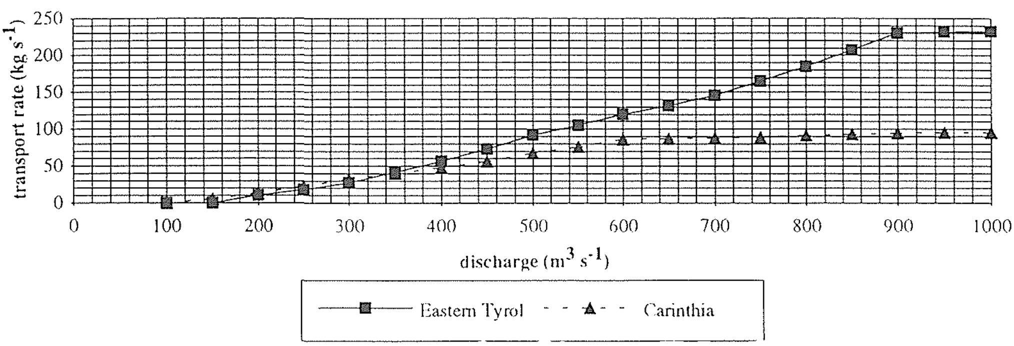

Fig. 3 Bed load transport rates at two different sites for the River Drau.

Temporal variability

The reason for taking 1931 to 1991 as a reference time period, in order to look at spatial variability of bed level changes, is that the largest engineering measures were already established by 1931, and afterwards stable conditions would have been ideal from an engineering point of view. In order to determine detailed temporal developments of the bed changes we have split the whole period into three subperiods (Fig. 4). The period 1931-1960 was characterized by tendencies for aggradation in reaches 1, 2, 3 and 5, whereas in reach 4 degradation occurred. From 1960 to 1969 reach 2 was especially affected by a large degradation rate of 1.5 m maximum. Most of the bed underwent degradation in reaches 1, 3, 4, 5. The larger degradation in reach 2 was caused by further regulation measures to protect a village after the catastrophic floods of 1965/1966.

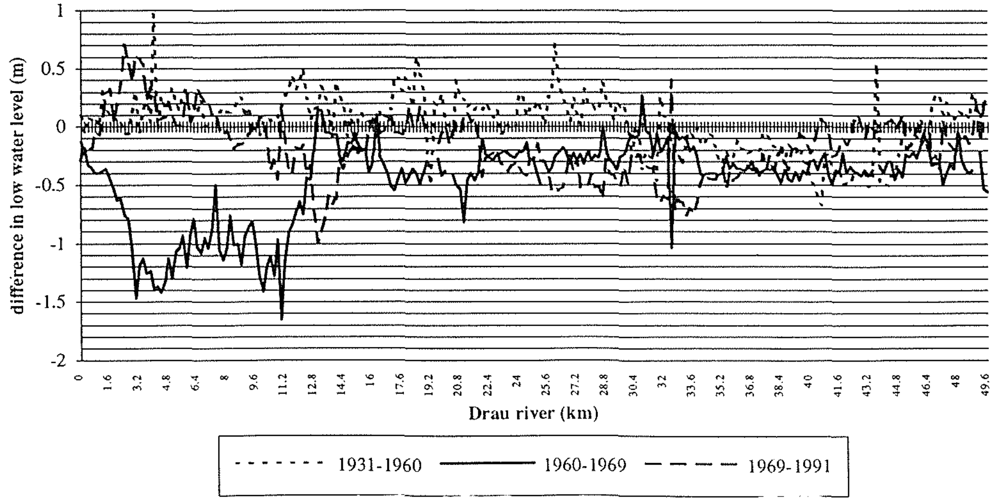

Fig. 4 Changes of low water level for different time periods.

By comparison 1969-1991 shows nearly opposite conditions in reach 2, with aggradation rates almost 0.6 m, whereas reach 3 was again more stable; reach 4 degraded up to 0.7 m and reach 5 slightly degraded. It is important to stress, that viewing just a certain period of time (for example 1969-1991) may lead to incorrect interpretations and conclusions with regard to the need for future engineering measures.

It is interesting to note, that since 1886 mean bed level changes for short periods of time show a rapid change between aggradation and degradation (Fig. 5). For the period 1886-1906 the mean value of bed level change was more than 0.15 m over the whole 50 km reach. Such aggradation was typical for this time period, as the whole reach was dominated by aggradation, before regulation measures were realized. Until 1914

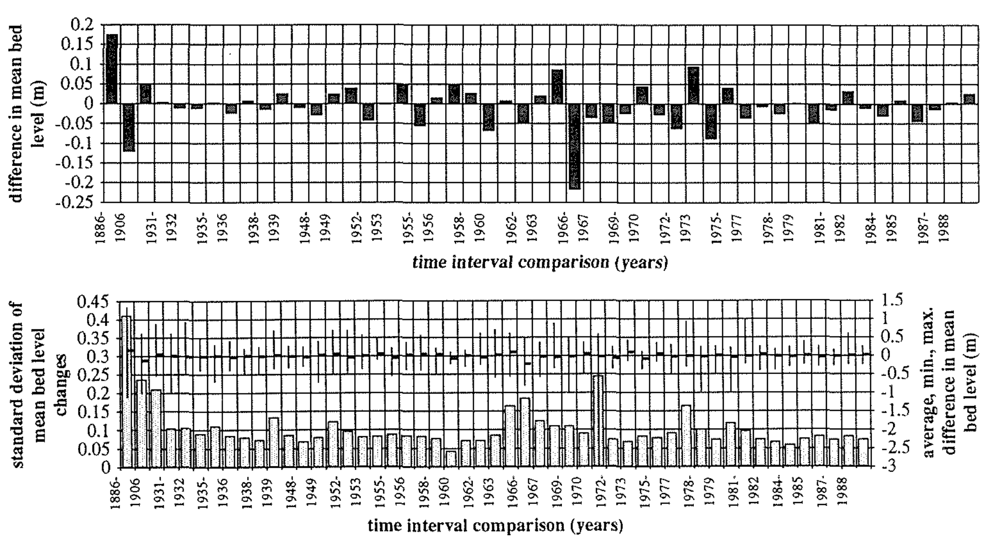

Fig. 5 Average bed level changes for time interval comparison within a time period from 1886 to 1993.

degradation was dominant. This had lead to the conclusion, that the engineering measures were successful.

Following this there were periods with aggradation as well as degradation but in general, degradation was dominant. The years 1965-1967 appear to have had the greatest influence for the period which followed. The big floods of 1965/1966, with 100 year discharges led to aggradation, immediately followed by considerable degradation from 1967. This prompted further regulation measures and gravel mining. During *c.* 15 years there was a situation comparable with the 1930s, perhaps with higher degradation rates. Minimum and maximum differences in low water level are now mostly within ± 0.5 m, whereas formerly they were much larger. The very high standard deviation at the beginning of the century can be explained through the aggrading system that, with engineering measures, was transformed into a slowly degrading one in a relatively short period of time. Higher degradation rates were typical because of the introduction of engineering measures after 1965/1966.

When planning engineering measures for a management project it is crucial to consider the temporal and spatial variability in detail in order to evaluate river complexity. The time dependant development of special locations along the whole reach of the River Drau can be seen in Fig. 6. Some of the graphs are interrupted because of missing data for some years. Whereas at km 0.4, the difference between 1964 and 1969 is not significantly high, other locations downstream had degradation rates of almost 1 m. Downstream of km 16, this dramatic change cannot be observed and also the long-term degradation rate is not significant. Regression analyses show significantly high degradation rates over a period of more than one hundred years further downstream of km 16 and upstream of km 45 (Fig. 7). The lowest locations also show degradation, but not so significantly.

The reasons for the different behaviour of these river reaches might be found in the varying mean diameters of subsurface material, sediment input through small tributaries

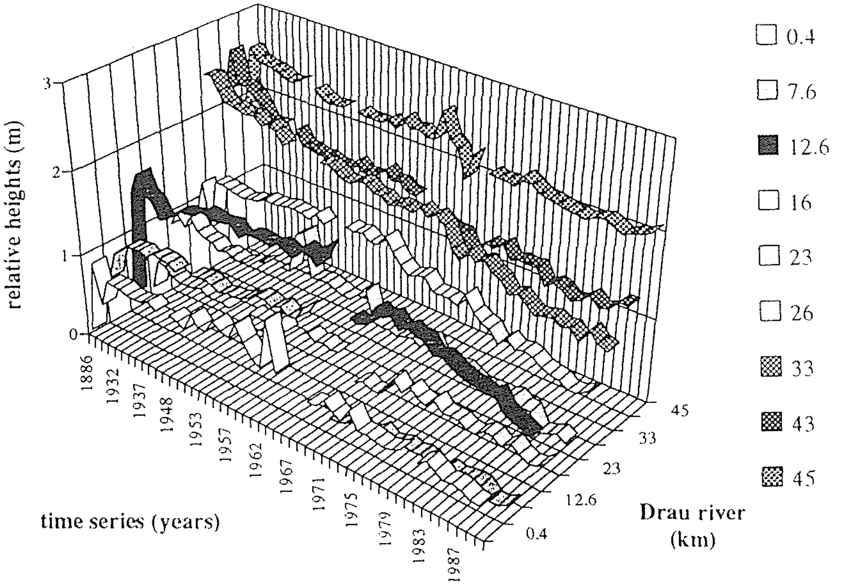

Fig. 6 Temporal and spatial variability of low water level developments at selected locations of the River Drau.

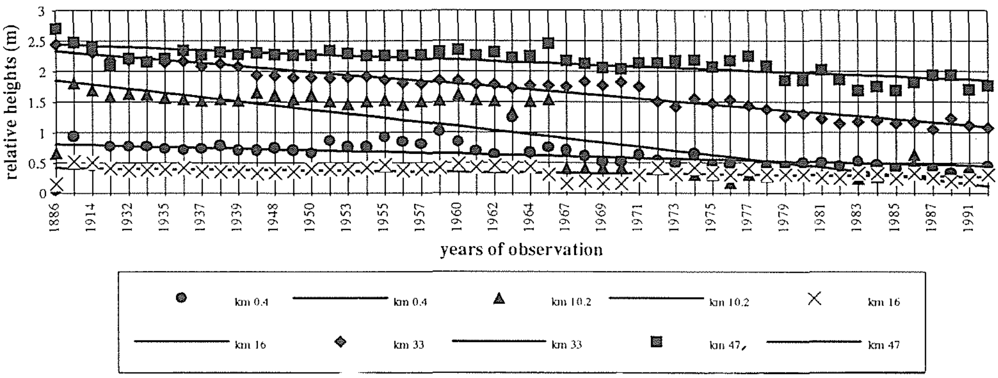

Fig. 7 Regression analysis of long-term river bed changes at different locations.

along the whole reach, selective transport, grain sorting, armouring effects and gravel mining. Selective transport is determined for the River Drau by downstream fining. Abrasion might be negligible, as Parker (1992) showed for quartz one of the main rocks transported by the tributary Isel. One important reason might be that a large amount of the transported sediments is deposited at the upper part of the reach and, therefore, further downstream there is a lack of material. This can also be noted in Fig.6, where aggradation takes place for the aggradation period 1886-1906 only until km 16, whereas downstream degradation occurs except for the last few kilometres.

CONCLUSIONS

The River Drau in Austria has problems related to the instability of bank protection measures and ecological deficits. The analysis of the data for a 50 km long reach of this river leads to the following conclusions:

(a) There is a definite relationship between sediment transport and human impacts.
(b) The uniform design of the cross sections is not appropriate to maintain a stable dynamic river bed.
(c) The volumetric amount of degradation corresponds to 60% of gravel abstraction.
(d) The input from small tributaries is important for the longitudinal changes in the river bed. Sediment input from these tributaries into the main river is mainly caused by local rainfall processes. Thus short term data describe local phenomena.
(e) To apply simulation models to rivers with variable beds, long-term observations are required and initial results indicate that such models can be applied for the River Drau.
(f) Short term data sets cannot reveal changes in of river morphology over long distances.
(g) Low water level measurements appear to be a cheap and fast method for monitoring bed level changes especially for long river reaches.
(h) Many closely spaced measuring points are more important for the analysis of sediment transport developments in relation to human impacts than only a few, daily measured ones.

(i) Detailed longterm monitoring programmes, including the whole drainage basin are necessary to explain sediment transport variability thereby allowing sound engineering design. A working group was established to plan and perform a future monitoring programme for the River Drau.

(j) The complexity of managing such rivers is determined by the large temporal and spatial variability of sediment transport.

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Land-use effects on magnitude-frequency characteristics of storm sediment yields: some New Zealand examples

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Abstract Magnitude-frequency relationships for storm suspended sediment yield were analysed at four small basins under various land uses near Auckland, New Zealand. Relationships between storm sediment yield and storm peak flow rate were combined with continuous records of streamflow to derive (a) the return periods of given storm sediment yields, (b) the long-term average sediment yield, (c) the proportion of the long-term yield carried by storms of various return period. The results showed that an urbanizing basin yielded much more sediment with larger rarer storm events, and the bulk of its long-term average sediment yield was carried by events with longer than annual return periods. In contrast, in a nearby mature urban basin storm yields increased little with larger storm flows, and it was the weekly-monthly events that carried the most sediment in the long term. For pasture and market-gardening basins, the sub-annual and super-annual events were of approximately equal importance in carrying the long-term yield. These patterns of storm sediment yields probably relate mainly to landuse controls on erosion processes and sediment supplies. The results suggest different approaches for the most cost-effective controlling of basin sediment exports.

INTRODUCTION

Greater public interest and improved legislation for resource management in New Zealand over recent years has increased the need for knowledge of stream sediment loads and sediment exports from small basins. Traditionally, measurements of basin sediment exports have focussed primarily on obtaining an estimate of the long-term average (or mean annual) yield. Often, in climatic regimes containing considerable interannual variability, the time constraint for obtaining planning information conflicts with the length of time required to obtain sediment yield data that are representative, and results in quite uncertain estimates of the "long-term average" sediment yield. Also, while a basin sediment yield value is useful to manage sediment problems downstream, such as sediment supplies to a reservoir, or to suggest how basin exports might change after a land-use conversion, it is less useful where there is a need to manage sediment routing within a basin, since in-basin sediment problems tend to arise on an event basis.

Sediment yields from individual storms and their magnitude-frequency characteristics are rarely assessed, yet they have the potential to enable better estimates of the long-term basin yield and to identify the return periods of the events that transport the most sediment in the long-term. For example, is more sediment exported in the long term by a few large rare storms or by the relatively small but frequent events that occur monthly to several times yearly? Such knowledge could then be used to design the most cost-effective means of sediment management.

This paper presents results from analysis of individual storm sediment yields in four small basins under various land uses but experiencing the same climatic regime. The purposes are to present an alternative method for estimating a long-term representative basin sediment yield, to show how the magnitude-frequency characteristics of the storm sediment yields differ, to discuss the possible reasons for this, and to explore some implications for in-basin sediment management.

DATA AND ANALYSIS

Study basins

Data are used from four basins in the greater Auckland area (Table 1). Pasture, mature urban, developing urban, and market-gardening land uses are represented. Basin size is small, ranging from 0.3 to 2.2 km^2.

Table 1 Study basin characteristics, flow records, and sediment yield estimates.

Basin	Land use	Area (km^2)	Flow records (year)	Storms sampled	Storm rating regression coefficient	Yield by storm rating (t km^{-2} $year^{-1}$)	Yield by frequency analysis (t km^{-2} $year^{-1}$)	T_{50}* (year)
Alexandra	Urbanizing	2.2	1.2	22	0.97	960	2370	3.0
Wairau	Mature urban	1.5	6	17	0.79	107	100	0.2
Whangapouri	Market gardening	1.8	3	34	0.78	49	52	0.4
Manukau	Pasture	0.3	17	14	0.92	49	46	0.8

* T_{50} Return period at which 50% of long-term sediment yield is moved by storms of shorter return period.

Data collection

Data collection procedures were essentially the same at all basins. At each basin outlet, streamflow was monitored by continuously recording water levels at a "V-notch" weir. A nearby auto-sampler was triggered to collect samples of water from storm flows for determination of suspended sediment concentration. The samplers became activated when the water level exceeded a threshold value, then continued to sample through-out the storm event at either fixed intervals of time or changes in water level. The records of suspended sediment concentration and water discharge collected in this fashion were then combined to calculate the discharge of suspended sediment over the storm event. At Alexandra basin, depth-integrated samples were collected over a range of discharges and sediment concentrations in order to establish a relationship between the point concentration at the auto-sampler intake and the mean cross-section concentration.

There, the mean/point concentration ratio was found to be 0.95 ± 0.05 and this was applied during the calculation of storm yield. At the other basins, a 1:1 relationship between point and mean concentration was assumed. Automatic tipping-bucket-type rain gauges were operated within the basins to record storm rainfalls.

Analysis methods

Determining sediment yields In practice, because of malfunctions and the limited number of sample bottles with the auto-samplers, not all storms could be sampled for suspended load. Accordingly, the strategy was to measure accurately the sediment yield from a sampling of storms and from these determine relationships between storm sediment yield and storm peak flow, with these relationships then being used to estimate the yields in unmeasured events over the complete flow record. Least-squares regression was used to parameterize these relationships. When calculating the sediment yield over the period of flow record, storms were distinguished on the basis of discrete quickflow events. The yields from very small events were ignored if the event quickflow was less than a threshold value. Separate threshold quickflows were identified for each basin.

A previous study by Hicks (1990), in which storm sediment yields at a number of small New Zealand basins were related to total runoff during the storm, quickflow runoff, peak discharge, Williams' (1975) runoff factor (i.e. the runoff * peak flow product), storm rainfall, rainfall energy, and Wischmeier & Smith's (1958) EI_{30} rain erosivity factor, showed that storm sediment yields correlated best overall simply with the storm peak discharge.

Magnitude-frequency relationships Analysis of storm event frequencies involved first compiling a peaks-over-threshold series of all peaks in the flow record. This series was then transformed to one of storm sediment yields by applying the storm sediment yield ratings described above. After assigning return periods to the elements of the storm yield series (using the Gringorten formula), an Extreme Value model (type 2) was fitted to the magnitude-frequency distribution. The goodness-of-fit was tested using the method of Hosking *et al.* (1985). Working with the EV model over a recurrence interval band from 0.1 to 20 years, the 20-year average annual yield transported by events in 0.1

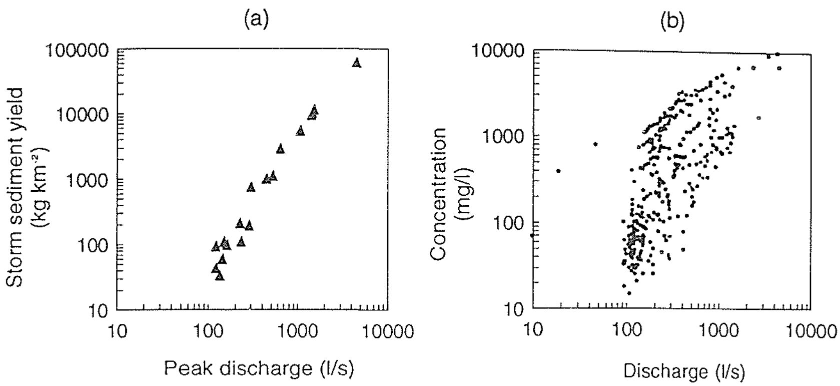

Fig. 1 Relationships between (a) storm sediment yield and peak flow and (b) instantaneous sediment concentration and water discharge for Alexandra site.

year increments of return period was determined by multiplying the probability-of-occurrence of events in that increment by the storm yield corresponding to the mid-point of the return period increment. Summing the yields over all increments provided an estimate of the 20-year average annual sediment yield.

The 20-year time base used in this study was considered to be an acceptably long period to represent the "long-term" yet did not require excessive extrapolation of the magnitude frequency relationship. Any arbitrary time base could be used with the approach.

RESULTS AND DISCUSSION

Storm yield *vs.* peak flow

The storm sediment yield *vs.* peak discharge "ratings" were "crisp", with high regression coefficients for the logarithmically transformed data (Table 1). At all four basins, the statistics of the storm yield ratings were superior to those for instantaneous sediment concentration and water discharge ratings. A typical storm sediment yield rating and instantaneous concentration rating are compared in Fig. 1. It appeared that while the instantaneous concentration could vary by up to a factor of one hundred for a given discharge at most basins, due to apparently random supplies of sediment to the stream, this "noise" tended to be averaged-out over the storm event.

Magnitude-frequency characteristics

The storm sediment yield magnitude-frequency relationships varied among the study basins (Fig. 2), with the largest variations appearing to be related to land use. The overall position of the magnitude-frequency relationships on Fig. 2 is indicative of the relative sediment yields from the basins, while the steepness of the lines indicates the relative yields between frequent and rare events.

The relationship for the urbanizing Alexandra basin (28% bare ground, 18% scrub and trees, 52% pasture, and 2% urban at the time of the study) shows that, compared to the other basins, it yields more sediment from storms over all return periods. Moreover, its steepness shows that the sediment yield increases exponentially as the storm magnitude increases, with extremely high yields predicted for rare events. In

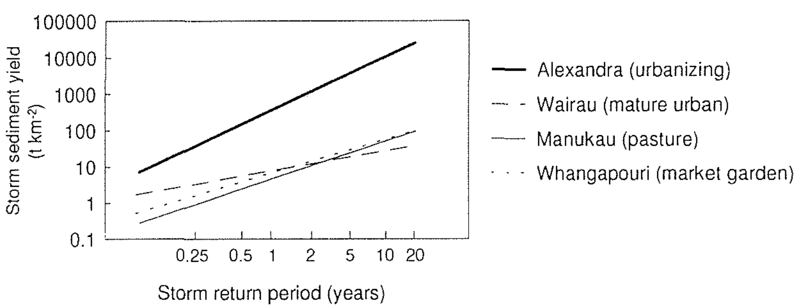

Fig. 2 Magnitude-frequency relationships for storm sediment yields.

comparison, the relationship for the "mature" totally urban Wairau basin is relatively flat, indicating that not a great deal more sediment is yielded from large, rare runoff events than is yielded by events that occur several times per year. This behaviour can be related to the ground cover. The large degree of paving in the urban basin limits sediment sources, and the sediment supply is probably exhausted early during the larger storms. Conversely, the supply of sediment from exposed soil slopes in the urbanizing basin is relatively unrestricted, with gullying and sheet erosion processes expected to intensify during larger events.

The "crossover" of the relationships for the urban Wairau basin and the pasture basin at Manukau shows that the urban basin yields more sediment during events occurring more frequently than 2-yearly, but the pasture basin yields more from less frequent events. Again, this is consistent with a sediment-source/erosion-process effect, with slope erosion only becoming relatively more important in the pasture basin during bigger, rarer events, while sediment supplies in the urban basin appear to be limited for all but the most frequent events.

The market-gardening Whangapouri basin responds in a similar fashion to the pasture basin. The relatively low yields from the market-gardening basin were surprising in light of the expected high erosion rates in the tilled fields, and point to a very low sediment delivery ratio.

Figure 3 shows the cumulative proportions of the 20-year average sediment yields carried by events with return periods less than the plotted value for the four basins. It is derived from the relationships in Fig. 2 weighted by the cumulative probability of occurrence of events in any one year. It shows that in the urbanizing Alexandra basin, the 20-year average sediment yield is carried relatively equally by events of short and long return period – for example, the 20-year event does a similar amount of work as does the annual event. This contrasts with the other three basins, where the most effective events at exporting sediment in the long term are those that occur several times per year. In the mature urban Wairau basin, it is the weekly-monthly events that carry the most sediment in the long term.

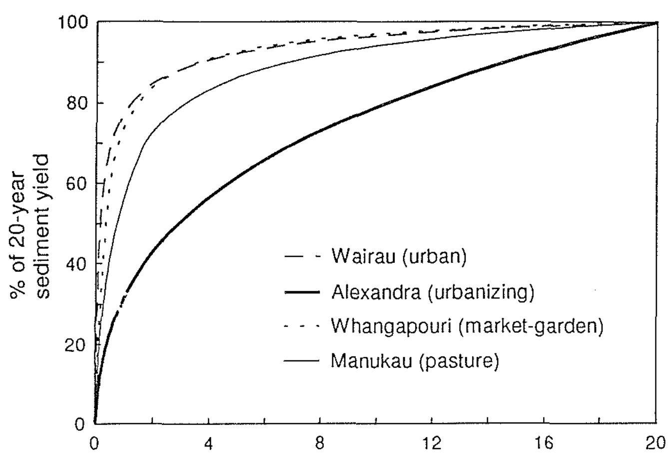

Fig. 3 Cumulative proportions of 20-year average sediment yield carried by storms of given return period.

A useful index of the relative geomorphic importance of shorter and longer return period events was found to be the return period at which events of that recurrence interval and less carried 50% of the long-term (i.e. 20-year) yield. The values of this "T_{50}" index for the four basins are compared in Table 1. A T_{50} of 0.2 years (2-3 months) for the mature urban basin clearly indicates the predominance of the frequent events, while a T_{50} of 3.0 years for the urbanizing basin points to the much greater importance of the rarer events. For both the rural basins, the sub-annual and super-annual events are of more equal importance. Again, as discussed above, these results appear to relate to the influence of land use on erosion processes and sediment availability.

Long-term average yield estimates

The "long-term average" basin sediment yields estimated by both the "storm yield rating" approach and by using the storm yield magnitude-frequency approach are compared in Table 1. The agreement between the two approaches varies. For example, the 20-year time-base figure for Alexandra basin (2370 t km^{-2} year^{-1}) is more than a factor of two higher than the yield estimated simply by using the storm yield rating over the one year of record (960 t km^{-2} year^{-1}). While partly due to the uncertainties in modelling the storm yield ratings and the magnitude-frequency relationships, the differences in the results from the two methods also highlight the uncertainty that can occur when the "long-term" yield of a basin is based on a record for only one or two years.

For basins with a "steep" magnitude-frequency relationship, such as Alexandra, the actual yield during any short (one or two year) period will be very dependent on what events occur then. Even for less responsive basins, the random occurrence of storms can lead to considerable inter-annual variability in sediment yields. An idea of how variable the sediment yield can be from year to year in the Auckland region is indicated in Fig. 4, which shows annual yields and also 3- and 5-year running means over the 16 years of record at the pasture basin at Manukau. The annual variability ranges over a factor of 10, while even the 5-year mean varies by a factor of two.

These figures indicate that considerable caution should be used when comparing estimates of mean annual sediment yield among basins where the yields are based only on short non-concurrent periods of record.

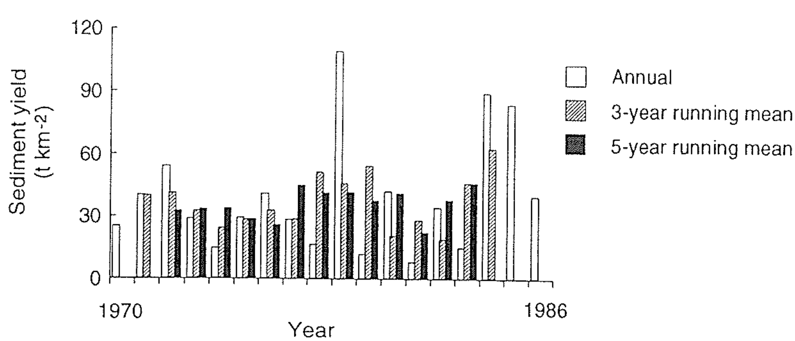

Fig. 4 Annual sediment yields at Manukau site, plus 3- and 5-year running means.

Generally, if it can be assumed that basin land use remains stable, the yield determined with the magnitude-frequency approach provides a fairer means of comparing yields among basins with flow records of varying length, since the estimate is based on a standard time base for all basins (i.e. 20-years for this study) and incorporates the contributions from events of return period longer than the record period. For cases where land use is changing, such as with the urbanizing basin, the yield determined by the magnitude-frequency approach is only a potential yield keyed to the condition of the basin during the period of record. In reality, as a basin is urbanized, its storm yield magnitude-frequency response will change, and so its actual long-term average yield will be less than if it had been left in its disturbed state.

Implications for basin sediment management

A clear conclusion from this frequency analysis is that the long-term average sediment yields from the basins in established land use are not excessively influenced by extreme events. Conversely, the yield from a basin whose soils have been temporarily exposed for urban development and are easily eroded will depend very much on the magnitude of the largest event that occurs before the ground is covered again. This contrast suggests different approaches for in-basin sediment management. Thus for basins with stable land use, particularly the urban ones, the best return would likely be achieved by targeting the very frequent sub-yearly storms. For urbanizing basins, the rare events should be targeted when designing sediment control structures and basin development should be scheduled to minimise the probability of an extreme event occurring while the basin remains vulnerable to erosion. Thus development work should be done quickly, and areas with exposed soil should not be left for extended periods. Possibly also, ground clearing could be restricted to certain seasons if a seasonal affinity for extreme events could be shown.

CONCLUSIONS

Conclusions from the study are:

(a) The study basins showed good relationships between storm sediment yield and storm peak flow.

(b) The urbanizing basin yields much more sediment with larger rarer storm events than do the basins under stable land use. The mature urban basin yields little more sediment as storm events get larger. These different sediment yield responses to different return period events probably relate to differences in sediment supplies and erosion processes among the basins. In particular, while sediment supplies probably become exhausted during large storms in the mature urban basin, sediment supplies appear to increase dramatically during larger events in the urbanizing basin due to intensifying gully and sheet erosion.

(c) In the urbanizing basin, the bulk of the long-term average sediment yield is carried by events with longer than annual return periods. For both the pasture and market-gardening basins, the sub-annual and super-annual events are of approximately equal importance. In the mature urban basin, it is the weekly-monthly events that carry the most sediment in the long term.

(d) There can be considerable uncertainty in estimates of basin sediment yield when the yield is averaged over only a few years of record. Estimating the average annual yield over a standard time base, such as 20 years, using the storm-yield magnitude-frequency relationship provides a fairer means of comparing sediment yields among basins with stable land use.

(e) For controlling sediment yields from basins with stable land use, the greatest benefit would probably be derived from targeting the relatively frequent, small events. In urbanizing basins, the largest events should to be targeted and the ground surface should be uncovered for as short a time as possible in order to minimise the probability of large events occurring and causing catastrophic erosion.

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Suspended sediment from two small upland drainage basins: using variability as an indicator of change

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Abstract Ten years of sediment sampling in the Balquhidder drainage basins, Scotland, has shown that variability of the suspended sediment concentrations dominates the data sets. The two basins are physically similar in all aspects except land use; the Kirkton being recently clearfelled and the Monachyle afforested. Comparison of the variability in the annual data sets with land use, climatic and hydrological changes indicate that in the Kirkton basin variability increased when the clearfelling started but variability in the Monachyle did not increase until a year after the pre-afforestation ploughing. Recovery of both basins was slowed by changes in the rainfall and runoff patterns. Standard deviations of flow determined sub-sets of the sediment data were used to compare the data from the three periods in the record: pre-disturbance, disturbance and recovery. The Kirkton showed an increase in the variability in high flows during the second and third periods but no change in low flows. The Monachyle showed an increase in variability at low flows during the second period but an increase in variability at high flows and decrease in low flows in the third period. The differences between the basins are thought to be due to the different sediment sources, with those in the Kirkton being more hillslope sources, whereas those in the Monachyle are in-channel sources. It is concluded that variability is an important component of the sediment regime and can be used to indicate changes in basins.

INTRODUCTION

Ten years of sediment sampling in two upland basins in Scotland has shown that the variability of suspended sediment concentrations dominates the data sets. The sampling was part of a water quality programme within a larger research study to determine the impacts of forest management on water resources. Early attempts to quantify the impacts of land use changes in the basins were hampered by excessive scatter in the data; severe smoothing seemed to be the only way of developing a time series, but as the data set increased the scatter emerged as being an important component of the data. Analysis of the scatter has now been carried out and systematic variability found which can be used to demonstrate the impact of changes in the basins.

Previous results from basin experiments rarely discuss the variability of the sediment data, but Guy (1965) noted that the process of erosion causes a highly variable quantity of sediment to be suspended, Walling & Webb (1987) gave an examples of excessive variability, where there was a lack of any relationship between concentration and streamflow, and Olive & Rieger (1992) reviewed a range of studies to show how natural variability in streams affects research results and how this can only be resolved by increasing the length of record.

The suspended sediment data sets for the two basins in Scotland indicate that changes have occurred during the period of sampling, but it has been difficult to quantify those changes and to identify the causes. The initial approach was to use a rating curve method (Johnson, 1988) with correction factors applied to take into account the scatter in the data (Ferguson, 1986). Temporal changes in suspended sediment-flow relationships were identified for the first few years, but scatter in the data became excessive making the further use of rating curves dubious. During the period of the experiment there were significant variations in the climate, with rainfall and flow increasing. These variations complicate the main objective of the sediment programme, which was to determine the effects of the land use changes.

This paper aims to look at variability of the suspended sediment concentrations in the Balquhidder basins, Scotland, to see if it can be used to indicate the impacts of the land use and climatic changes.

CHANGES IN THE BALQUHIDDER BASINS

The Balquhidder paired basin experiment was established in 1981 to study the effects of forest management on water resources (Johnson & Whitehead, 1993). The basins are located some 60 km north of Glasgow in the southern Highlands of Scotland. They are physically similar, some 7 km^2 in area with an altitude range from 250 to 900 m. The geology is mica schist and soils a mixture of ranker, surface water gleys and peaty gleys. The major difference between the basins is the land use, with the Kirkton basin, at the start of the experiment, supporting a 42% cover of coniferous forest and the Monachyle basin rough grazing.

In early 1986 the land uses in both basins were changed with clearfelling of the Kirkton forest and afforestation of the lower altitudes in the Monachyle. Clearfelling included the removal of tree branches on site, with the timber extracted to roads by skylines, then uplifted by articulated lorries. Afforestation included the pre-planting ploughing for drainage of the wetter ground, but leaving 15-30 m wide buffer strips in the riparian zones. Both land use changes affected relatively small areas of the hydrological basins, but this is typical of upland forestry in the UK. Table 1 includes figures for the annual areas clearfelled and afforested. Sediment sampling started in late 1982, so three years of baseline data were available before the land uses changed.

During the period of the experiment there were significant variations in the rainfall and streamflow patterns. Cumulative daily rainfall showed a significant increase in late 1986 and the mean daily rainfall intensity and frequency of 24-h rainfall totals exceeding a threshold of 30 mm (a large event in Scottish terms) both showed considerable variations in the period (Table 1). Changes were observed in the hydrological regimes of the basins, with cumulative daily runoff increasing and both mean daily runoff and the frequency of flood events in excess of 20 mm day^{-1} varying (Table 1).

The basins have therefore been subjected to changes in the land use, rainfall and runoff patterns which could potentially have had an impact on the sediment release and transport in the fluvial systems. By developing a catalogue of changes (Table 1), it was hoped to relate changes in the suspended sediments to specific events within the basins.

Table 1 Tabulation of changes in sediment variability, land use, rainfall and river flow in the Balquhidder basins for the sediment years 1983-1984 to 1991-1992.

Change	Sediment year: 1983-84	1984-85	1985-86	1986-87	1987-88	1988-89	1989-90	1990-91	1991-92
1	5.5	4.9	9.5	10.2	12.8	12.8	12.9	13.7	14.9
2	108	45	125	367	62	499	500	207	244
3	10.0	10.7	10.3	10.1	14.6	13.0	12.0	12.3	12.2
4	22	38	39	51	132	85	186	100	116
5	0	0	4	60	19	25	17	0	0
6	0	0	40	5	0	0	0	0	0
7	8.3	9.0	10.9	8.5	9.1	9.8	11.1	9.3	10.3
8	199	193	186	223	237	232	215	211	222
9	10	8	20	9	11	12	19	16	15
10	4.6	4.3	6.7	5.0	4.7	6.7	6.4	4.8	-
11	9	6	22	16	9	20	21	15	-

Changes:
1 Area of parallelogram (Kirkton);
2 Standard deviation of suspended sediment concentrations (Kirkton), mg l^{-1};
3 Area of parallelogram (Monachyle);
4 Standard deviation of suspended sediment concentrations (Monachyle), mg l^{-1};
5 Kirkton area of clear felling, ha;
6 Monachyle area of ploughing, ha;
7 Mean daily rainfall intensity, mm;
8 Number of rain days;
9 Number of days with rainfall > 30 mm;
10 Mean daily runoff (mm);
11 Number of days with runoff > 20 mm.

CHANGES IN THE SUSPENDED SEDIMENT REGIMES

The 1983-1992 time series of suspended sediment concentrations for both basins (Fig. 1) show that the data are dominated by the variability and it is very difficult to determine any changes or trends in the mean values. Some 4000 samples were taken in each basin over the 10 year period, which is likely to provide a sample large enough to represent the true distribution (Johnson, 1992). Suspended sediment concentrations are generally greater in the forested Kirkton basin compared to the Monachyle basin, the 10-year mean concentrations being 114.7 and 43.7 mg l^{-1}, respectively. In terms of the scatter of points, the standard deviations of the data are also higher for the Kirkton compared to the Monachyle, 352.3 and 117.6 mg l^{-1}, respectively.

Due to the large variability in the data, envelopes were constructed around the data to indicate the extreme values throughout the data set. This was carried out manually on a stepwise basis for both basins, independently and with no consideration of basin or

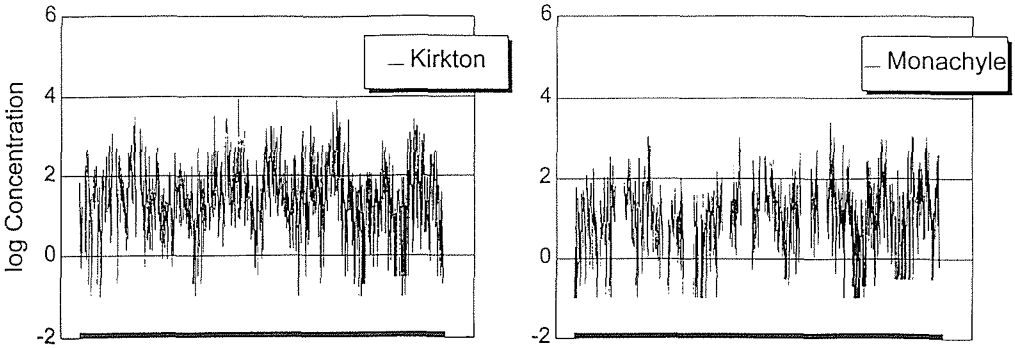

Fig. 1 Time series of the suspended sediment samples for the Balquhidder basins, 1983-1992 (concentration in mg l^{-1}).

climatic changes. When the envelope changed appreciably due to the addition of a further group of data then this was noted as a break in the time series and a new envelope started. This can be somewhat subjective and needs an honest approach to the analysis. The results were that four periods were identified for the Kirkton basin and five for the Monachyle basin, with the breaks between the periods occurring around the time of a summer low flow period. This result suggested that the sediment regimes of the rivers were reacting on an annual time cycle, starting and ending at the time of a prolonged dry period. Consequently the data for both basins was subdivided into sediment years stating on 1 June and finishing on 31 May.

Quantification of the sediment envelopes was difficult, so each one was simplified by drawing parallelograms around the data, fixed in the horizontal by flow limits and manually drawing a straight line which best encompassed the extremes of the data. Two of the sediment years (1984-1985 and 1991-1992) are shown in Fig. 2. The areas of each parallelogram were calculated from the sediment concentration which each corner represented (Table 1).

The results of the parallelogram analysis show that area, and hence the scatter of the extreme values, in the Kirkton increased in the years when felling occurred, but in the later years, when the felling temporarily stopped, the scatter increased again. In the Monachyle, scatter increased over a year after the ploughing was carried out then decreased, but remained at a higher level than during the pre-disturbance years.

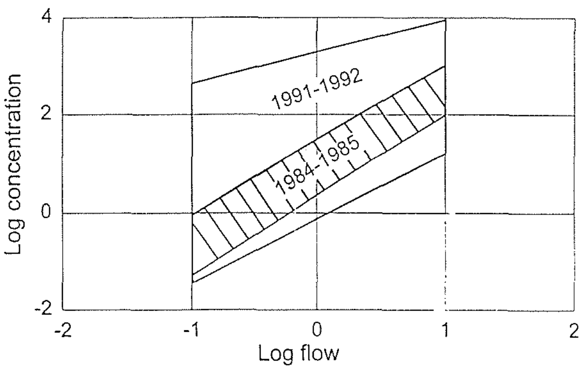

Fig. 2 Examples of the parallelogram analyses for the Kirkton basin containing the extreme data points in the two years 1984-1985 and 1991-1992.

VARIABILITY, SCATTER AND LAND USE CHANGE

Variability is a rather loose term which has been used here to indicate scatter. For these sediment data, a statistical term was sought to represent variability so that consecutive sediment years and the basins could be compared. Various options were available but it was considered most appropriate to select one widely used term and not broaden the analysis to find terms which may misrepresent the data; standard deviation was the term selected. For each sediment year a number of analyses using different data sub-groups, such as low flows and high flows, were tried, details of which can not be included here due to lack of space.

Table 1 is a selection of the sediment year calculations for the period 1983-1984 to 1991-1992 and provides a summary of the most relevant results. To represent the variability in the data, both the area of the parallelograms and the standard deviation of the annual data are presented to be compared with results for land use changes, climatic and hydrological variations.

A comparison of the data in Table 1 shows that there was no simple and consistent relationship between the sediment concentration variability and the basin data. The first three years of the study contained a mixture of climatic years with 1983-1984 and 1984-1985 having few storms, but 1985-1986 having many more, the storms in the third year mostly occurring in the months prior to the land-use changes. There appears to be no reaction in the Monachyle sediment regime to the climatic variability before the land uses changed but the Kirkton showed an increase in variability in 1985-1986, even though the clear felling did not start until after the storms occurred. This suggests that the Monachyle basin had a natural protection against more extreme climatic years, but the Kirkton had less natural protection possibly due to the past afforestation.

The parallelograms showed a two phase increase in the Kirkton, but a single peak in the Monachyle. The first changes in the Kirkton parallelogram occurred over the period 1985-1988, initially associated with the frequent storms but later with the clearfelling. The second increase in 1990-1992 can not be related to any new land-use changes as no clearfelling occurred in these years. There was, however, an increase in the mean daily rainfall intensity in this period, and although other climatic factors changed, none appears to provide a good explanation for the second increase in the parallelogram area. The Monachyle showed an increase in the parallelogram area in 1987-1988 over a year after the ploughing, then a decrease for the last three years, but not back to the background levels. It is likely that the one year delay was due to the ploughing occurring in a relatively storm free year, but once the next year's storms had fully activated the plough lines the following years remained more variable than before the disturbance.

The standard deviation of the data in each sediment year for both basins again show increases which can be related to the land-use changes, with the Kirkton reacting immediately and the Monachyle being delayed by over a year. The Kirkton results are in general agreement with previous work, for example Burgess *et al.* (1981) who found that there was an immediate reaction to logging operations in Australia. The Monachyle results differ slightly from other results, for example Robinson & Blyth (1982), who found an immediate reaction to pre afforestation ploughing. Peaks in the Kirkton standard deviation occurred in 1986-1987, 1988-1989 and 1989-1990, the first of which can be related to the land use but others more closely to the number of large storm and runoff events. The Monachyle standard deviation peaked in 1987-1988, due to the land use, but again in 1989-1990, again probably due to the storm events.

The above analysis can only be indicative of the cause and effects in the sediment regimes, but with land use, climate and hydrology all changing it remains extremely difficult to separate the effects.

VARIABILITY AS AN INDICATOR OF CHANGE

The previous section identified three contrasting periods in the Kirkton and Monachyle basins over the 10 years of monitoring: a 2 (Kirkton) and 4 (Monachyle) year baseline period followed by a 3 (Kirkton) and 1 (Monachyle) year period of reaction to land uses and finally a 4 year period in both basins of recovery from land-use changes, but reaction to climatic changes. For comparison between the two basins these three periods have been standardized to the years 1983-1985 (period 1), 1985-1988 (period 2) and 1988-1992 (period 3). For each period the suspended sediment data were subdivided into groups according to the stream flow at the time of the sample.

Results of the group analysis showed that, in general, there was an increase in the standard deviation with flow, with the Kirkton having a closer relationship than the Monachyle. Results for the three periods were very different (Fig. 3): the standard deviations at high flows in the Kirkton increased in successive periods, but at low flows

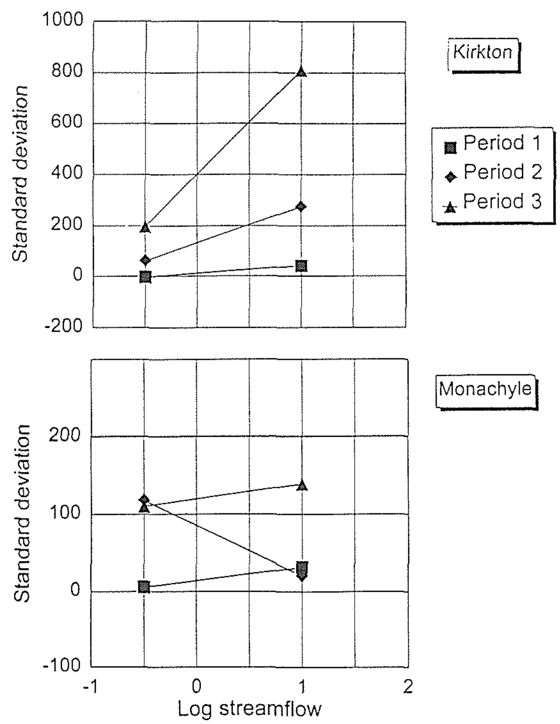

Fig. 3 Regression lines of the relationships between standard deviation of the suspended sediment samples and streamflow for the periods pre-disturbance (period 1), disturbance (period 2) and recovery (period 3).

showed no change, while in the Monachyle the second period showed an increase in standard deviation at low flows, but little change in high flows resulting in a reversed relationship with flow. In the third period the relationship reverted back to being parallel to, but shifted above, period 1. The above results differ from the parallelogram analysis, in that the later showed an increased scatter in all flow ranges; this is the contrast between looking at the whole sample in the flow range with standard deviation, but only the extreme values with the parallelograms.

Comparing the standard deviations for the two basins the first period showed relatively small differences compared to the subsequent two periods. In periods 2 and 3, the main impact of basin changes in the Kirkton were confined to the high flow samples, but in the Monachyle the high flow samples were only affected in period 3, and low flow samples only in period 2. The change in the relationship and the contrast between the basins was substantial and it is considered that the cause was due to the land use changes creating new, but different, sediment sources. The main new source in the Kirkton was on the clearfelled hillslopes, which are only linked to the river system during storm events, while the new Monachyle sources were within the drainage system where a limited supply of material was removed in the lower flows, leaving little available for transport in the higher flows.

CONCLUSIONS

The main results of this work are:

(a) The forested basin had the greater suspended sediment variability before the land uses changed, as the moorland basin appeared to be naturally protected against storm events.

(b) The standard deviation of the samples and the envelopes containing the extreme data points showed a progressive change in response to both the land-use and climatic changes.

(c) The land-use changes increased the variability in the data, at high flows in the Kirkton and low flows in the Monachyle. This was seen as the land use creating new, but different, sources of sediment in each basin.

(d) Climatic extremes after the land use changes had a greater impact than similar events before. This was due to the land use changes destabilizing the basins, possibly for a very long time.

(e) Variability can be used as an indicator of change, but long data sets are required.

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