

Sediment supply from landslide-dominated catchments: implications for basin-margin fans

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ABSTRACT

The sediment flux from a mountainous catchment can be expressed as a function of a landslide rate constant κ which accounts for the vigour of hillslope erosion. Since the incising drainage network flushes all or a portion of the products of hillslope erosion to a range front where fan deposition takes place, a conservation of solid sediment volume allows the fan area and progradation distance to be calculated. These parameters are related primarily to the discharge of sediment from the catchment and to local tectonic subsidence.

A survey of modern alluvial fans in a wide range of climatic and tectonic settings shows that the effects of climate and bedrock lithology cannot be discriminated in the scatter of data of catchment area vs. fan area. However, by focusing on over 100 fans in the arid and semiarid zone of SW USA, the impact of tectonic subsidence rate is unambiguous. Although further quantitative data on local tectonic subsidence rates are urgently required, our preliminary analysis suggests considerable potential for reconstructing palaeocatchments where basin tectonic subsidence rates can be estimated. The progradation distances of fans from the northern and southern margins of the Middle Devonian Hornelen Basin of Norway, and the western and north-eastern margins of the Mio-Pliocene Ridge Basin, California, allow catchment sizes and denudation rates to be approximated. Although unique solution sets are not possible, an iteration of parameter values allows plausible parameter combinations to be calculated which shed light on the tectonic and sedimentary history of the proximal basin and upland source regions. Model results suggest significant asymmetry in basin subsidence rates, catchment slopes and transport mechanics between the two margins.

INTRODUCTION

It is clear from modelling studies (e.g. Steckler *et al.*, 1993; Heller *et al.*, 1993; Burgess & Allen, 1996; Rivenaes, 1997) that the sediment discharge into a basin is a first-order control on sequence stratigraphic architecture, acting as a boundary condition for far-field dispersal. The magnitude of the sediment discharge, its calibre and its spacing along the range front are primary parameters controlling the gross geometries and depositional facies of sedimentary bodies, such as alluvial fans, along the proximal edge of the basin (Paola *et al.*, 1992; Heller & Paola, 1992; Gordon & Heller, 1993; Whipple & Trayler, 1996). Consequently, an improved understanding of how catchment-delivery systems function, preferably over long (that is, geological) time-scales and in different tectonic and climatic settings is of fundamental importance to the modelling of basin stratigraphy.

The sediment production rate in a drainage basin, or

sediment yield, has been studied over a wide range of spatial and temporal scales. Many investigations have tended to focus on individual drainage basins where bedrock lithology, local climate, land use practices and morphometric properties are well-constrained. Others have amounted to a global mapping of the sediment yields of a large number of river drainage basins derived from suspended sediment load data obtained at downstream gauging stations (e.g. Milliman & Meade, 1983; Milliman & Syvitski, 1992; Summerfield & Hulton, 1994; Hovius, in press a) with some attempt at evaluating the controlling factors. Recent studies have exploited the strong coupling between tectonic uplift/subsidence and erosion in process-based models involving drainage network evolution and hillslope processes (Willgoose *et al.*, 1991; Kooi & Beaumont, 1996; Tucker & Slingerland, 1996; Hovius *et al.*, 1997; Densmore *et al.*, in press). Such process-based studies provide important insights which may be of use in formulating general models of sediment flux at geological time scales.

The most obvious setting where a sediment supply control on the development of sedimentary geometries

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and sequences can be evaluated is the deposition of alluvial fans at mountain range fronts. Firstly, in many cases the catchment-fan is a closed system amenable to a balance of mass or volume. Second, the progradation distance of fans, because of their simple geometry, is a length scale readily observable in the stratigraphic record which can be directly related to the sediment supply and pattern of subsidence.

The aim of this contribution is therefore to present a simple theoretical model of alluvial fan progradation with application to geological contexts. We model the sediment delivery from the catchment as a function of the rate of landsliding. Using a conservation of solid sediment volume, the modelled sediment discharge is then distributed into a basin with a simple pattern of tectonic subsidence to obtain typical alluvial fan progradation distances which can be compared with ancient geological examples. Following the lead of Gordon & Heller (1993) and Whipple & Trayler (1996), our analysis sheds further light on the importance of tectonic subsidence to the relationship between drainage basin area and fan area originally proposed by Bull (1962, 1964), and offers new insights into the interpretation of alluvial fans at tectonically active mountain fronts in the geological record.

SEDIMENT EFFLUX OF MOUNTAINOUS CATCHMENTS

Mountain belts are narrow zones of tectonic convergence in which crustal material is processed rapidly due to high rates of erosion. Two process systems combine to degrade the regional topography resulting from crustal shortening (Hovius, in press b). One is responsible for valley lowering. The other ensures removal of mass from the adjacent hillslopes. Valley lowering is effected principally by fluvial incision, although glacial processes play a role at higher latitudes and at high altitudes near the equator. Once valley systems have developed, their higher order elements concentrate so much drainage and sediment transport that efficient down-cutting is likely under all but the most arid conditions. Such drainage elements, often orientated at high angles to the structural grain of the orogen, become entrenched in the uplifting bedrock and are thought to maintain their vertical position over geological time intervals (e.g. Burbank *et al.*, 1996). Between valleys, bedrock is uplifted towards denuding hillslopes, whose rate of mass transfer is limited by the rate of bedrock weathering. If this rate is slower than the local rate of valley lowering, then the hillslopes continue to steepen, becoming increasingly susceptible to landslides involving bedrock. Thus, rock-strength-controlled landsliding effectively puts an upper limit on the amplitude of local relief, and time-independent topography ensues where this mechanism is dominant. In such areas, long-term average sediment yield is solely determined by rock uplift rates, and is, as a consequence, independent of climate (Whipple & Trayler, 1996; cf. Pinter & Brandon, 1997).

Humid, rapidly uplifting mountain belts, such as the Southern Alps of New Zealand, the Central Range of Taiwan and the Karakorum of the western Himalayas, are believed to have steady-state, landslide-dominated topography (Suppe, 1984; Adams, 1985; Burbank *et al.*, 1996). In the western Southern Alps, the flux of material due to landsliding is roughly equivalent to the sediment output of principal rivers draining the region (Hovius *et al.*, 1997). The sediment efflux of such mountainous catchments is controlled at first order by the spatial and temporal patterns of slope failure. In a general sense, these patterns can be captured in a magnitude-frequency analysis of slope failures. Within a geologically homogeneous region, the landslide magnitude-frequency distribution follows a robust power law (Hovius *et al.*, 1997). This distribution may be written in a cumulative form

$$n_c(A \geq A_c) = \kappa (A_c/A_r)^{-\beta} A_r \quad (1)$$

where $n_c(A \geq A_c)$ is the number of slides per year of magnitude greater than or equal to A_c over a reference area A_r , κ is the rate of landsliding per unit area per year and β is a dimensionless scaling exponent equivalent to the slope of the power law. It implies scale invariance (Turcotte, 1992) of landslides across a range of length scales, and integration over the entire scale range of the process yields estimates of long-term hillslope erosion. Differences in rock properties between mountain belts may cause the scaling exponent of the power law to vary, inducing differences in the relative importance of high-magnitude events (Hovius *et al.*, in review).

The timing and short-term intensity of hillslope mass wasting are strongly controlled by the probability distribution of trigger events. Although landslides ultimately result from the oversteepening of slopes or slope segments, the vast majority of slope failures coincide with episodes of either prolonged or heavy rainfall, or seismic activity. Landslides therefore occur in a spatially and temporally clustered fashion, the rate of slope failure strongly depending on the intensity of the triggering event. Thus, the long-term average sediment yield of a mountainous region is made up of prolonged episodes of subdued or zero hillslope mass wasting and short intervals of enhanced erosion. In the eastern Central Range of Taiwan, where recent landsliding has primarily been climate-triggered, daily erosion rates may be over two orders of magnitude greater than the decadal average (Hovius *et al.*, in review). Although such high rates are not sustained for periods longer than a few days, their effect can surpass the average annual sediment yield of a catchment. More extreme still are erosion rates achieved during high-magnitude seismic events in some regions. For instance, a seismic cycle, including two M 7.1 earthquakes, lasting for three months in late 1993, set off over 4200 landslides in the western Finisterre Mountains, Papua New Guinea. Only several hundred slope failures had occurred in the same region during the decade preceding this episode, implying a co-seismic increase of

erosion rates by three orders of magnitude, sustained for several months (Hovius, unpublished data).

The interface of the hillslope and valley floor systems is characterized by a discontinuity of transport mechanics. Sediment efflux of a mountainous catchment is therefore as much constrained by the transport capacity of the stream network into which the debris is transferred as by the mode and rate of hillslope mass wasting (Montgomery & Buffington, 1997; Hovius *et al.*, in review). The short-term dual supply and transport control on downstream sediment load is most pronounced in mountainous drainages where valley lowering occurs principally through fluvial incision. In such cases, the resulting valley geometry has little accommodation space at the base of hillslopes, forcing a direct transfer of hillslope mass wasting products into the fluvial system. Subsequent removal of all but the coarsest products of hillslope mass wasting occurs within days of its mobilization, unless a discrepancy exists between the amount of sediment supplied and the transport capacity of the river at the site of entry. Such discrepancies may arise, for instance, from arid conditions with reduced water discharge, or from the release of large amounts of material during high-magnitude seismic trigger events, and can result in the damming of streams. Failure of landslide dams (Costa & Schuster, 1988) is often accompanied by extremely high water and sediment discharges, with profound effects on foreland sedimentation. Further complications may result from intramontane storage of hillslope mass wasting products, where valley geometry permits this. Important build-ups of talus and colluvium may form where only part of the valley floor is occupied by an active river system, as is the case in many glacially widened valleys. Subsequent erosion of these deposits adds to the downstream sediment load a component not directly related to contemporaneous hillslope mass wasting. Thus, intramontane storage acts to decouple the sediment efflux of a mountainous catchment from the ultimate source area of that sediment on the hillslopes. The cumulative effect of these complicating factors is not to propagate the scale invariance of landslides into the downstream sediment loads of mountainous catchments.

In the analysis that follows we extend our concepts from mountainous terrains in convergent settings to the smaller scale ranges associated with crustal extension and apply the resulting model to geological examples with well-developed basin margin facies. The theoretical model is a development from that published by Whipple & Trayler (1996). It differs in the parameterization of the sediment supply from the eroding catchment and in a minor way in the use of efficiency factors for sediment routing through the catchment-fan system. Our model incorporates a uniform distribution of subsidence in the basin, in contrast to the spatially variable patterns examined by Whipple & Trayler (1996). Such spatially variable patterns of subsidence, which have been shown to exert a strong influence on fan dimensions, could be relatively easily bolted on to our model.

THEORETICAL MODEL

Since the volumetric discharge of sediment from a mountainous catchment can be estimated from the rate of landsliding, this information can be used in a theoretical model of deposition in a basin located immediately against a range front where the landslide discharge provides a boundary condition for sediment dispersal and deposition in the basin (Fig. 1).

Although the magnitude-frequency characteristics of the sediment load of rivers cannot be expected to correlate well with the scaling characteristics of landslides on the contributing mountainous hillslopes, nevertheless the time-averaged volumes of sediments discharged onto a neighbouring fan must be the same as the hillslope landslide discharge for a system involving no time-varying storage. Consequently, the landslide model can be used to make predictions about fan area and fan progradation, but is limited in its ability to predict bed thicknesses and volumes.

The theoretical model involves a source terrain (catchment) of area A_1 feeding a fan of area A_2 . The relation between fan area and catchment area has been reported from geomorphological studies of modern fans (e.g. Hooke, 1968; Bull, 1977; Lecce, 1990) as a power-law

$$A_2 = c A_1^n \quad (2)$$

where the exponent n and the coefficient c have values variously reported from the literature (Table 1) and attributed to factors such as variations in bedrock lithology, climate, rate of uplift of rock, rate and spatial distribution of subsidence. The reason for the correlation between fan area and catchment area is that as the latter increases the sediment discharge out of the catchment also increases. This discharge D_s is given by

$$D_s = \bar{\epsilon} (1 - \lambda_r) A_1 \quad (3)$$

where $\bar{\epsilon}$ is the spatially averaged erosion rate over the catchment (m yr^{-1}) and λ_r is the rock porosity in the catchment. The sediment discharge has units of $\text{m}^3 \text{yr}^{-1}$. It is important to discriminate it from the sediment yield or efflux which are normalized by the area of the catchment, with units of m yr^{-1} . The sediment discharge from the catchment due to landsliding is also given by

$$D_s \approx \frac{2\beta\varsigma\kappa}{(3-2\beta)} L_1^{(3-2\beta)} (1 - \lambda_r) A_1 \quad (4)$$

derived from Hovius *et al.* (1997), where ς is a width-depth scaling relation for landslide scars (taken as 0.05 ± 0.02 by Hovius *et al.*, 1997), β and κ are taken from Eq. (1), and L_1 is the upper length scale for landslides. β appears to be a constant related to the intrinsic properties of the landslide process and falls in the region 0.7–1.16 (Fuyii, 1969; Sugai *et al.*, 1994; Hovius *et al.*, 1997). The range of values may be related to the anisotropy of rocks in the catchment (Densmore *et al.*, 1997). The maximum length scale for landsliding is set by the dimensions of the hillslopes available, approximating half the drainage spacing where valley sides are not strongly dissected (Fig. 1). A value of 1 km

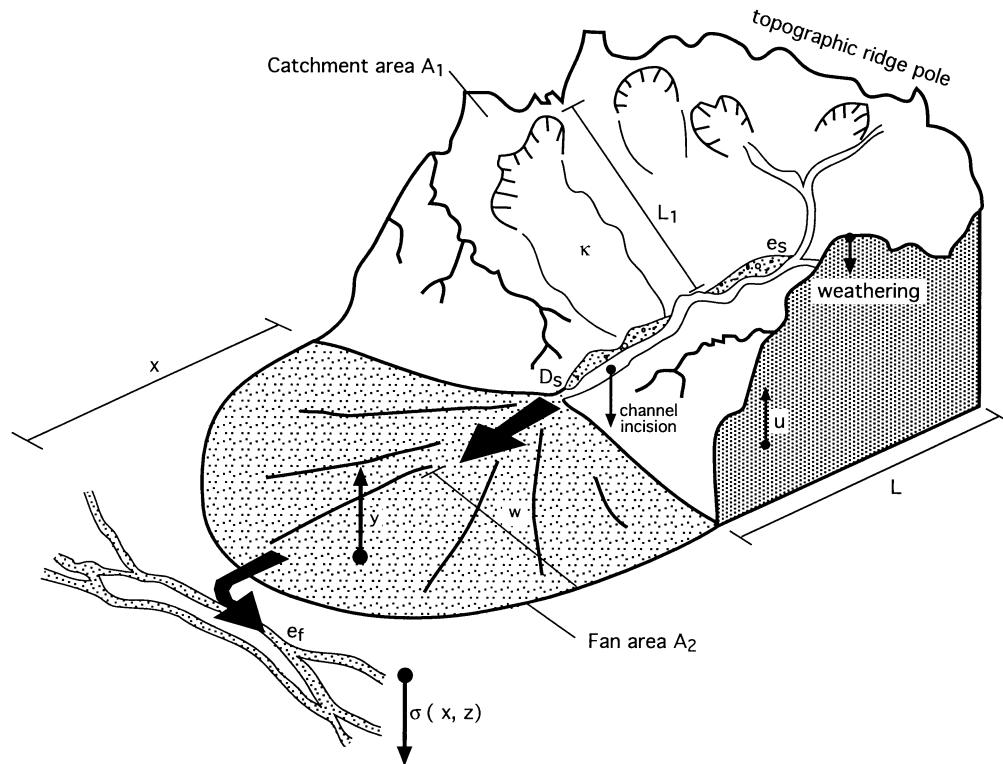


Fig. 1. Schematic diagram of a landslide-dominated hillslope system feeding an alluvial fan to illustrate main parameters and notation used in the text.

was considered appropriate in the Southern Alps of New Zealand. Consequently, the sediment discharge due to landsliding can be seen to be determined primarily by the rate constant κ . In the tectonically extremely active Southern Alps of New Zealand, the value of κ derived from a 60-year record of landslides is about $5 \times 10^{-5} \text{ km}^{-2} \text{ yr}^{-1}$ (Hovius *et al.*, 1997). We can therefore write a simplified version of Eq. (4) as

$$D_s \approx V(1 - \lambda_r)A_1 \quad (5)$$

where V , the sediment efflux, is primarily a function of κ . When the landscape is in steady state, the channel incision rate approximates the tectonic uplift rate of rock, as emphasized by Whipple & Trayler (1996). Since V is equivalent to the denudation rate due to landsliding, it can be viewed as a surrogate for the tectonic uplift rate where landsliding dominates hillslope erosion.

Some material, derived by hillslope erosion, may be stored in valley bottoms instead of being transported immediately beyond the range front onto the fan surface. The efficiency of sediment transport from hillslopes to fan apex therefore affects the volume of material delivered to the fan. To account for this, we introduce an efficiency factor e_s where the subscript refers to storage. The storage efficiency factor varies from zero (complete storage) to 1 (no storage). Hooke & Rohrer (1977) suggest that this storage efficiency factor is a function of catchment size, decreasing with catchment area, but empirical studies suggest the effect is relatively small (Hooke, 1968; Hooke & Rohrer, 1977; Jansson *et al.*, 1993).

We now make the assumption that the sediment discharge from the catchment is delivered to the fan and deposited in a closed system with no leakage to axial streams. If the fan is in steady state, sediment will accumulate up to a given profile, the accommodation space being provided by tectonic subsidence $\sigma(x, y)$. In a setting such as a simple graben with a planar border fault, the subsidence can be assumed to be spatially uniform in the directions x (normal to the fault) and z (parallel to the fault). The linkage between fan aggradation/degradation and local subsidence rate may be violated over short time periods when climatic or co-seismic effects may dominate, but over longer (geological) time periods this is a safe assumption. Fan size and rate of deposition clearly act in a trade-off, confirming the emerging view that tectonics, through the subsidence rate, is a first-order control on fan geometry.

There is the possibility, strong in humid fans, that some of the sediment delivered to the fan apex is transported through the fan and into down-fan river systems. Relaxing the assumption of a closed system to allow for such leakage, we introduce a second transport efficiency factor e_f (where the subscript refers to fan), which varies from zero (complete bypass) to 1 (entire sediment load deposited on fan surface) for net aggradation, but may be negative where fan erosion takes place. The final result for the fan deposition rate becomes

$$\dot{y} = \frac{1}{c} V \frac{(1 - \lambda_r)}{(1 - \lambda_s)} e_s e_f A_1^{1-n} \quad (6)$$

Table 1. Values of coefficient c and exponent n in the power law relationship between catchment area and fan area. The fans occur in a wide range of tectonic and climatic settings.

Region	c	n	Reference
<i>Arid region fans</i>			
Fresno County, California, shale catchments	2.4	0.88	Bull (1964)
Fresno County, California, sandstone catchments	1.3	0.88	Bull (1964)
Shadow Mountain, Death Valley, California	0.33	1.0	Denny (1965)
Winnemucca area, northern Nevada	0.74	0.98	Hawley & Wilson (1965)
Cactus Flats, southern California	0.24	1.01	Hooke (1968)
East side Death Valley, southern California	0.1	1.0	Denny (1965)
East side Death Valley, southern California	0.15	0.90	Hooke (1968)
East side Death Valley, southern California	0.16	0.90	Hooke & Rohrer (1977)
West side Death Valley, southern California	1.05	0.76	Hooke (1968)
West side Death Valley, southern California	0.95	0.89	Hooke & Rohrer (1977)
Little Cowhorn Valley, southern California	0.20	1.07	Hooke & Rohrer (1977)
Deep Spring Valley, southern California, dolomite/quartzite catchment	0.44	0.62	Hooke (1968)
Deep Spring Valley, southern California, quartzite catchment	0.16	0.75	Hooke (1968)
Little Cowhorn Valley, southern California	0.28	1.13	Hooke (1968)
Owens Valley, southern California	0.42	0.94	Hooke (1968)
South-western Bare Mountain, Nevada	2.05	0.70	Ferrill <i>et al.</i> (1996)
Eastern Bare Mountain, Nevada (excluding Tarantula Canyon fan)	0.55	0.92	Ferrill <i>et al.</i> (1996)
Northern sector, White Mountains, California	0.01	2.82	Whipple & Trayler (1996)
Southern sector, White Mountains, California	1.66	0.34	Whipple & Trayler (1996)
Western White Mountains, California and Nevada, erodible catchments	0.65	0.65	Lecce (1991)
Western White Mountains, California and Nevada, resistant catchments	0.70	1.22	Lecce (1991)
Sierra Nevada western bajada, Owens Valley, California	2.59	0.64	Whipple & Trayler (1996)
Owens Lake, southern California	0.64	0.47	Whipple & Trayler (1996)
Elburz Mountains, northern Iran	1.18	0.95	Beaumont (1972)
Ventura, California (a)	3.84	0.55	Rockwell <i>et al.</i> (1984)
Ventura, California (b)	0.59	0.8	Rockwell <i>et al.</i> (1984)
Spain	0.72	0.82	Harvey (1989)
Guadalentin Depression, Murcia, SW Spain	0.78	0.66	Silva <i>et al.</i> (1992)
<i>Humid and subhumid fans</i>			
Dellwood, North Carolina, USA	0.23	0.53	Mills (1982)
Roan Mountain, North Carolina, USA	0.38	0.76	Mills (1983)
General River Valley, Costa Rica	0.92	1.01	Kesel (1985)
Banff, Alberta, Canada, fluvial fans	0.48	0.32	Kostaschuk <i>et al.</i> (1986)
Banff, Alberta, Canada, debris-flow fans	0.17	0.48	Kostaschuk <i>et al.</i> (1986)
Japan (115 fans)	2.23	0.40	Oguchi & Ohmori (1994)
Polar region fans, Aklavik, NWT, Canada	2.38	0.65	Keeble (1971)

which can be solved for the fan area to give

$$A_2 = \frac{D_s}{ye_f(1-\lambda_s)} = \frac{1}{y} V \frac{(1-\lambda_r)}{(1-\lambda_s)} e_s e_f A_1. \quad (7)$$

This implies that the fan area is strongly dependent on the fan aggradation rate, itself determined by the rate of tectonic subsidence and the sediment supply.

In geological contexts, fan progradation is more easily observable than planwise fan area. We should expect a scaling relationship between the two. We assume that there is no headward growth of the catchment which acts as an internal feedback on fan area through our steady-state equations.

Let us now introduce a relation between fan progradation (x) and fan area

$$x = b A_2''' \quad (8)$$

which for a semicircular fan gives $b = \sqrt{(2/\pi)}$ and $m = \frac{1}{2}$, and for a half-elliptical fan with half-width w and length x gives $b = 2/\pi w$, $m = 1$. Since the fan area is given by Eq. (7), the progradation distance for the general half-elliptical case becomes

$$x = \frac{2}{\pi w y} V \frac{(1-\lambda_r)}{(1-\lambda_s)} e_s e_f A_1. \quad (9)$$

Equation (9) can be simplified by introducing a dimensionless parameter ϕ which incorporates the effects of distributing the sediment efflux to the fan head across a fan with a certain spatially averaged deposition rate:

$$\phi = \frac{1}{y} V \frac{(1-\lambda_r)}{(1-\lambda_s)} e_s e_f. \quad (10)$$

The dimensionless sediment dispersal parameter is a coefficient of proportionality between the progradation distance and the catchment area

$$x = \frac{2}{\pi w} \phi A_1 \quad (11)$$

and the progradation distance is by definition directly proportional to the fan area (eqn 8)

$$x = \frac{2}{\pi w} A_2. \quad (12)$$

Both Eqs (11) and (12) apply to the general elliptical case. If $\phi \ll 1$ gravels should be stacked very close to the basin margin against the range front. Alternatively, if $\phi \gg 1$ sediment will undergo far-field dispersal from the range front.

It may be helpful in some geological contexts to know the relation between the fan progradation distance and a simple length scale in the catchment. Hovius (1996) and Talling *et al.* (1997) have reported a relation between outlet spacing and the distance from range crest and range front, approximating catchment length L . For large transverse drainages in linear mountain belts Hovius (1996) found the aspect ratio of catchment length to width to be 2.2 ($L \approx 4.4w$). The aspect ratio is consider-

ably more variable in smaller catchments with variable bedrock geology, structural geometry and histories of network development (Talling *et al.*, 1997). The aspect ratio is particularly regular for fault blocks with linear range fronts and ridge crests. Talling *et al.* (1997) found the aspect ratio to vary between 1.41 and 4.06 for different fault blocks, with an average of 2.5. This is in strong contrast with the aspect ratios observed in flume experiments which show a range of 3.5–10.0 with a strong correlation with slope (Schumm *et al.*, 1987). It seems likely in natural drainage basins that the aspect ratio is established early in network development and may not correlate strongly with present-day slopes. It is therefore problematical to estimate reliably aspect ratios for fault block catchments in the geological record.

The relationship between catchment area and length of mainstream (L_s) (including the distance from the stream source to the drainage divide) is given by Hack (1957) for small catchment sizes:

$$L_s = 1.4 A_1^{0.6}. \quad (13)$$

Similar relationships have been proposed for a wider range of catchment sizes by Mueller (1972, 1973), and more recently by Montgomery & Dietrich (1992):

$$L_s = 1.78 A_1^{0.49}. \quad (14)$$

Mainstream length L_s can be related to catchment length L by stream sinuosity ($P = L_s/L$). Since mainstream sinuosities increase with catchment length and area (Smart & Surkan, 1967), there is a trend from relatively elongate small catchments to more equant larger catchments. The effect of sinuosity can either be absorbed in the coefficient of Hack's law, or, since P varies with A_1 , can be incorporated within the exponent. In steep mountainous catchments sinuosities are likely to be low. For the sake of transparency, we modify the coefficient in Eq. (14) slightly to incorporate the effects of sinuosity, which then becomes

$$L_s = 1.6 A_1^{0.49}. \quad (15)$$

The catchment length when combined with the outlet spacing defines the aspect ratio. This ratio, which has been measured in a wide variety of settings, serves as an additional constraint in the reconstruction of catchment sizes and dynamics from the geological record, which we term palaeocatchment studies.

RESULTS

The theoretical model expresses relationships between a number of parameters influencing erosion, sediment delivery to the range front and alluvial fan growth. These parameters encapsulate the effects of tectonics, climate and topography implicitly or explicitly. In order that fan deposition can be viewed in relation to one lumped parameter, rather than to a number of independently varying parameters, a dimensionless variable ϕ has been introduced (Eq. 10). The size of an alluvial fan relative

to its catchment area, and the progradation distance of a fan are both determined by ϕ . If the average denudation rate in the catchment approximates the average fan deposition rate and there is neither catchment storage nor fan bypassing, ϕ will differ from 1 only by a factor related to the different porosities of bedrock and fan sediment ($\approx 1.1\text{--}1.6$). Typical values of the dimensionless sediment dispersal parameter from over 100 modern fans are in the range $0.1 < \phi < 10$ (Fig. 2).

It is important to investigate the effects of the component parameters on the dimensionless grouping ϕ . These parameters are the tectonic subsidence rate or sediment deposition rate y' , the sediment flux from the catchment V expressed as a function of κ , and the efficiencies of storage or trapping in the catchment e_s , and of flushing through the fan to fluvial systems e_f . The rock and sediment porosities can be viewed as invariant for our purposes.

The distribution of tectonic subsidence close to the basin margin may be a simple uniform subsidence in the hangingwall of a border fault, a more complex three-dimensional pattern related to different amounts of slip along a fault segment, including rotation about a hinge-line, or a flexural response to crustal loading and unloading. These possibilities are discussed by Paola *et al.* (1992) and more recently by Whipple & Trayler (1996).

We consider only the case of a simple graben with

spatially uniform subsidence, where a uniform thickness of sediment is deposited on a semicircular fan surface. For a given catchment size, an increase in subsidence rate causes a reduction in ϕ and a decrease in fan area and progradation distance. At a value of the sediment efflux (V) of $10^{-4} \text{ m yr}^{-1}$ ($\kappa \approx 10^{-6} \text{ km}^{-2} \text{ yr}^{-1}$, where $\beta=1$, $L_1=1 \text{ km}$ and $\zeta=0.05$), tectonic subsidence rates in excess of 1 mm yr^{-1} are likely to promote stacking of fan gravels very close to the range front at values of $\phi < 0.1$.

The parameter κ contains within it a range of information on climatic impacts, bedrock and regolith characteristics, topographic slope and curvature effects and, in mountainous systems with rapidly incising streams, the dominating effects of rock uplift rates. The sediment discharge from the catchment is determined primarily by the landslide rate constant κ , but is also affected by the maximum length scale at which landsliding occurs and by the area-depth scaling parameter for landslides (Hovius *et al.*, 1997). Treating the latter two parameters as constant, fan area and progradation distance vary directly with κ and V . At a tectonic subsidence rate of 0.1 mm yr^{-1} , V is required to vary from about 10^{-3} to $10^{-5} \text{ m yr}^{-1}$ in order to stay within the range of $0.1 < \phi < 10$. The corresponding range for κ is $10^{-5}\text{--}10^{-7} \text{ km}^{-2} \text{ yr}^{-1}$ (where $\beta=1$, $L_1=1 \text{ km}$ and $\zeta=0.05$).

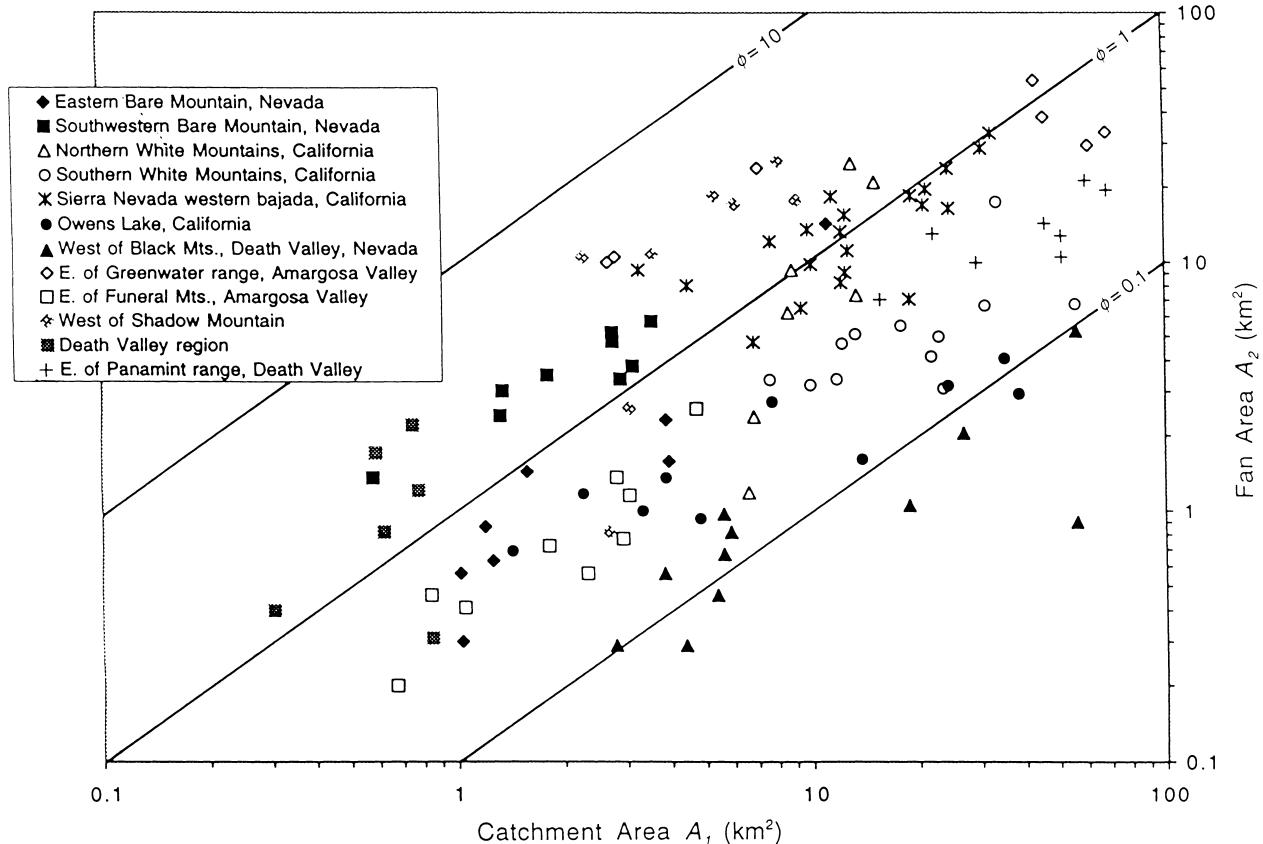


Fig. 2. Catchment area vs. fan area for 116 modern arid region fans, south-west USA. Data for Bare Mountain, Nevada, from Ferrill *et al.* (1996); White Mountains, Sierra Nevada and Owens Lake from Whipple & Trayler (1996); Death Valley and Amargosa Valley from Denny (1965). Most data fall between $0.1 < \phi < 10$.

The sediment efflux of the catchment is only the same as the hillslope denudation rate when there is no storage of sediment within the catchment. At long (geological) time-scales, catchment storage is unlikely in actively uplifting terrains. Low values of e_s imply high amounts of valley floor deposition, a reduction in hillslope gradients and a degeneration in the erosional system. This degenerative trend may occur after an instantaneous tectonic event uplifts the land surface but without continued tectonics during catchment development (Fraser & DeCelles, 1992). At shorter time-scales, substantial volumes of sediment may be temporarily stored in the catchment, the tendency for storage increasing with catchment size (Hooke & Rohrer, 1977) and the degree of drainage development (Schumm *et al.*, 1987). Fan area and progradation distance are directly proportional to the storage efficiency factor e_s .

Alluvial fans are both permanent repositories of sediment and temporary sites of storage of sediment *en route* for axial river systems. Fans may trap sediment delivered at the apex by hillslope erosion, but they may also be dissected. We concentrate on fan trapping here, where $0 < e_f \leq 1$, though we recognize that fan dissection is extremely important in estimating downstream sediment discharges. Fan area and fan progradation distance are both directly proportional to the fan efficiency factor e_f .

Catchment area vs. fan area

The relationship between catchment area and fan area has long been thought to be of fundamental importance (Bull, 1962, 1964), and the power law relation (eqn 2) has been examined, and confirmed, by a number of authors working in arid region fans (Denny, 1965; Hawley & Wilson, 1965; Hooke, 1968; Hooke & Rohrer, 1977; Lecce, 1991), humid region fans (Mills, 1982, 1983; Kesel, 1985; Kostaschuk *et al.*, 1986; Oguchi & Ohmori, 1994) and subarctic fans (Keeble, 1971). A comparison of coefficients and exponents in the power law relation between catchment and fan area for a wide range of climatic and tectonic settings shows a very wide variation (Table 1). There is nothing (in terms of n and c) that discriminates a fan in an arid region from a fan in a humid region, though it has been suggested that the slope of the power law relation differs significantly between arid and humid fans (Oguchi & Ohmori, 1994, p. 406). The slope of the power law, which in the great majority of cases is less than 1, indicates that as catchment size increases, the ratio of catchment to fan area also increases. This may be due to increased storage and lower sediment yields in larger catchments with lower mean slopes (Hooke, 1968; Church & Mark, 1980; Lecce, 1991), or to reduction in fan area by sediment bypass into downstream river systems (Hooke & Rohrer, 1977). The latter may be primarily responsible for the high exponents of some humid fans. The coefficient in the power law shows a wide variation unrelated to climatic setting. There is clearly another factor, or number of

factors, responsible for the value of the coefficient in the power law.

The plot of catchment area vs. fan area for 116 modern arid region fans in SW USA (Fig. 2) allows comparisons to be made between fans in different settings of subsidence rate. The seven fans of the northern White Mountains (Fig. 3) (Pakiser *et al.*, 1964; Whipple & Trayler, 1996) are derived from granitic bedrock and occupy a region of low tectonic subsidence in a simple graben. Their values of ϕ , obtained from A_2/A_1 (Eq. 7), range from 0.18 to 1.88. The nearby southern White Mountain fans form a narrow bajada and are derived from a mixed hinterland of granites, metasedimentary and metavolcanic rocks. The small fans are building out into a deep graben with inferred high tectonic subsidence rate (Pakiser *et al.*, 1964; Hollett *et al.*, 1989). Their values of ϕ are tightly clustered and range from 0.12 to 0.52 ($n = 12$). It is most likely, as suggested by Whipple & Trayler (1996), that the differences in tectonic subsidence rate between the northern and southern White Mountains are responsible for the marked variation in alluvial fan size along the range front.

A similar picture emerges from further south in Owens Valley (Fig. 3). The 21 fans along the western bajada of Owens Valley derived from the predominantly granitic rocks of the Sierra Nevada are building out onto a shallow bedrock bench with low subsidence rate. The values of ϕ range from 0.37 to 2.84 with a clustering around unity. Some of the source areas were glaciated during parts of the Pleistocene, while others remained unaffected by glacial processes, but these distinctive histories are not reflected in systematic differences of ϕ values. Further to the south along the same flank of Owens Valley (Fig. 3), fans derived from the granitic rocks of the Sierra Nevada are entering the deep Owens Lake graben with high subsidence rates. The values of ϕ from 10 fans range from 0.08 to 0.52, substantially lower than the values found along the western bajada to the north.

Within the central Basin and Range province, markedly different fan sizes and geometries are found on the eastern and south-western sides of Bare Mountain, Nevada (Ferrill *et al.*, 1996). Large coalesced alluvial fans form a bajada on the low-subsidence-rate south-western flank of Bare Mountain (Fig. 3). The nine fans studied by Ferrill *et al.* (1996) along this flank are sourced from Late Proterozoic and Palaeozoic clastic and carbonate sedimentary and metasedimentary rocks. The values of ϕ range from 1.16 to 2.33. In contrast, the Bare Mountain Fault causes rapid subsidence along the eastern flank, with subsidence rates increasing to the south (Reheis, 1988; Monsen *et al.*, 1992). Alluvial fans correspondingly decrease in area southwards, with values of ϕ ranging from 1.28 (northernmost fan) to 0.29 (southernmost fan). It is highly unlikely that factors other than tectonic subsidence rate can explain the clear trends around Bare Mountain.

One further example from Death Valley reinforces the

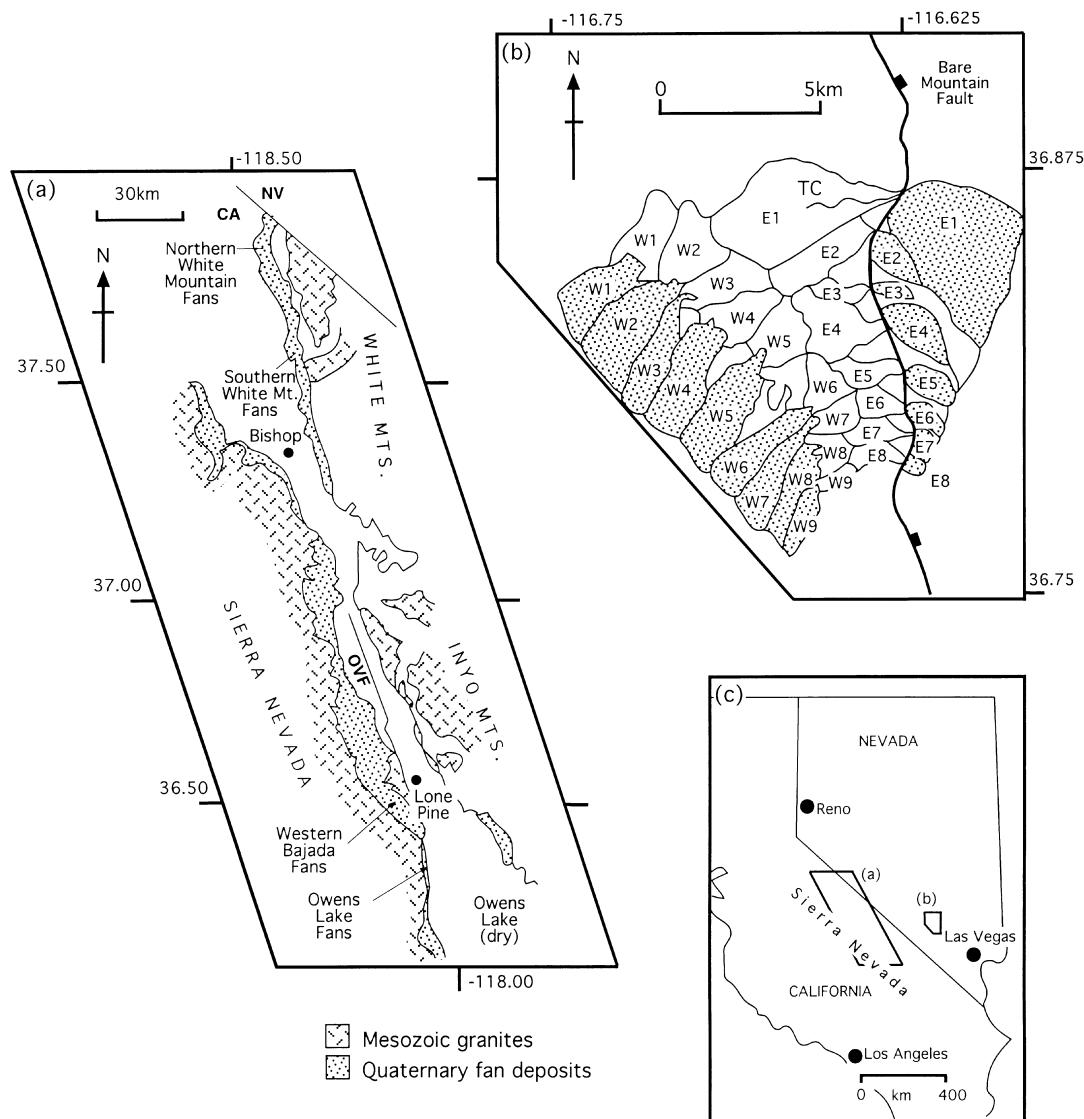


Fig. 3. Location maps of localities in the (a) Owens Valley – White Mountains area, California, and (b) Bare Mountain area, southern Nevada, showing catchments and fans on the eastern and western flanks. OVF, Owens Valley Fault; TC, Tarantula Canyon.

argument. Death Valley has developed as an extensional pull-apart basin along the dextral strike-slip Furnace Creek–Death Valley fault system. Tectonic subsidence is inferred to increase from west (Panamint Range) to east (Black Mountains) across the valley towards the Death Valley fault zone. Sedimentation rates, based on ^{14}C -AMS dating of the bulk organic fraction in recently recovered cores (Anderson & Wells, 1997) range up to about 1 mm yr^{-1} (typically $\approx 0.5 \text{ mm yr}^{-1}$) over a time period dating back to pluvial lake stages at $\approx 12 \text{ ka}$ and 40 ka . Morphological studies of fault scarps (Brogan *et al.*, 1991) suggest repeated slip during the late Quaternary, though no historical ruptures are definitely known. Along the Death Valley fault zone, fault scarps cut alluvium as young as late Holocene ($0.2\text{--}2 \text{ ka}$). On the opposite side of the valley, fault scarps are rounded and diffuse. The very young, unweathered fans on the east side of the valley derived from the Black Mountains

are small compared to the larger and partly weathered (varnished) coalesced fans derived from the Panamint Range (Denny, 1965) (Fig. 4). Although the suites of fans on opposite flanks of the valley are of slightly different age, they have each developed in response to denudation in their presently connected catchments. The values of ϕ are markedly different between the two fan populations (ranges of $0.02\text{--}0.17$ ($n=11$) for the east side, compared to $0.21\text{--}0.59$ ($n=8$) for the west).

Although the effects of climate and source area lithology may be locally important, our results point towards tectonic subsidence rate as the main factor responsible for the scatter of data in Fig. 2.

Fan progradation distance

Whereas catchment and fan areas can be easily measured in modern examples, this is impossible in geological

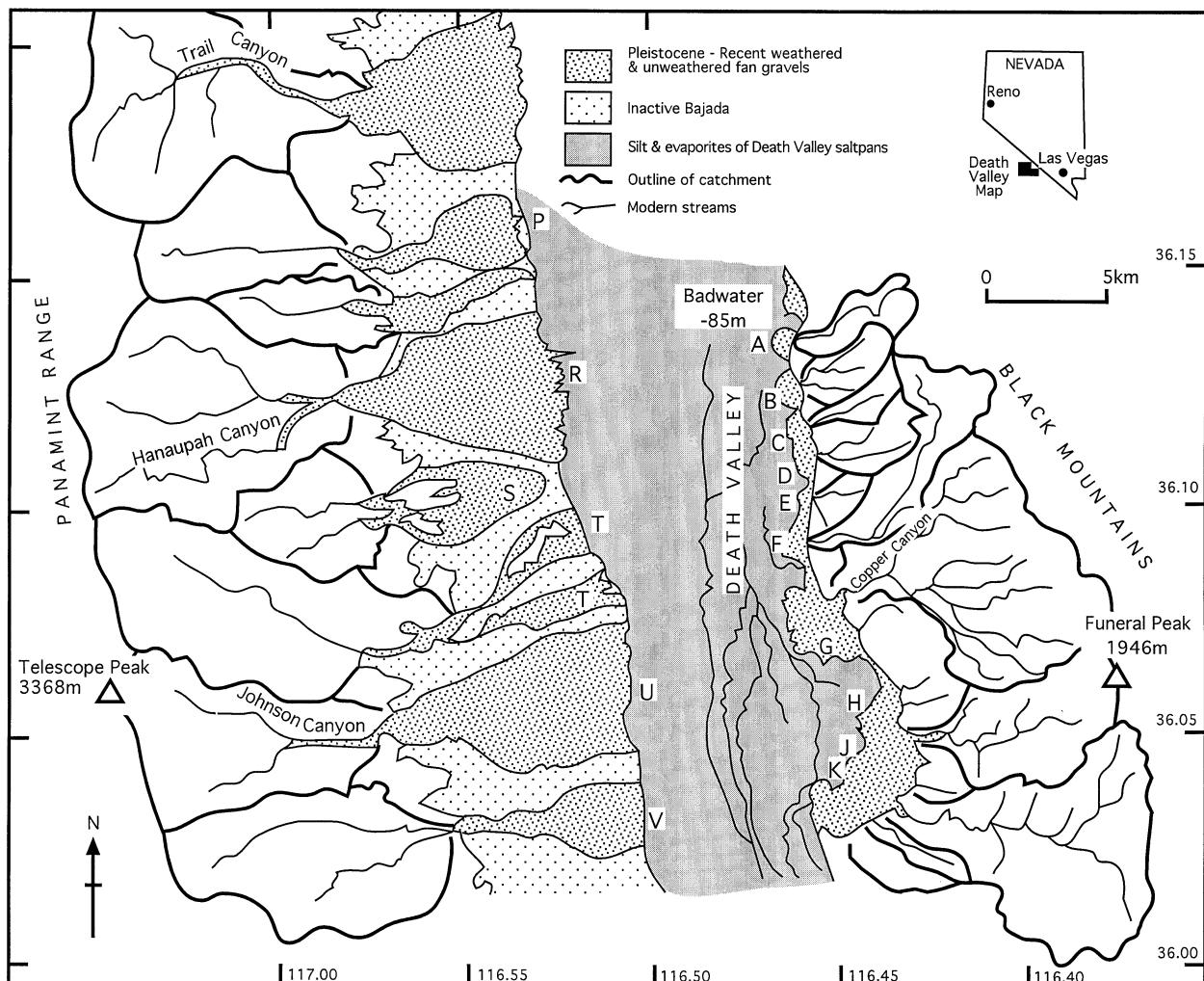


Fig. 4. Contrasting alluvial fans on either side of Death Valley (Denny, 1965) are explained by differences in tectonic subsidence rate. Fans labelled A to L and O to V refer to fans identified by Denny (1965), Plates 4 and 5.

situations where source regions have been eroded and fan bodies are incompletely exposed. Nevertheless, the basinward extent of the fan can be reconstructed from the distribution of sedimentary facies.

Consider a mountain range with a drainage network etched into it with a characteristic transverse spacing of the drainage outlets. During initial development of the drainage, or alternatively during steady state at very low values of ϕ , each fan will be separated from its neighbours. The basin margin is unsaturated with fan sediment. If x is the progradation distance and w the half-spacing of the outlets, the saturation factor s can be defined as $s = x/w < 1$. At higher values of ϕ a basin margin may reach saturation when fans are perfect planform semicircles that are just mutually touching ($s = x/w = 1$). At oversaturation fans are obliged to build basinwards as an apron or bajada, each fan body taking the form of an elongated ellipse ($s > 1$), corresponding to the elongation/elaboration phase of catchment development of Fraser & DeCelles (1992). The fact that semiarid and arid alluvial fans have length to width (aspect) ratios of generally less than 2 suggests that basinward progra-

dation quickly reaches a limit (the maximum extension stage of Fraser & DeCelles, 1992), after which the likelihood of sediment bypassing the fan and being exported into downstream fluvial systems increases. This export of sediment is reflected in the fan efficiency factor e_f . The percentage of sediment exported from perfect semicircular fans at the saturation condition depends primarily on γ and κ . Taking the eastern flank of Bare Mountain as an example of a laterally varying tectonic subsidence rate from ≈ 0.02 to 0.2 mm yr^{-1} , it can be seen that at the lower tectonic subsidence rate the fans would saturate and coalesce at very low sediment discharges from the catchment, corresponding to the coalesced fans near Tarantula Canyon at the northern end of the eastern flank of Bare Mountain (Fig. 3).

Palaeocatchment studies

Two examples are chosen to illustrate the application of the model to geological examples. In both examples, catchment sizes, average slopes and average denudation rates are estimated from a set of geological observables

comprising (a) progradation distance x (km), (b) fan aggradation rate \dot{y} (mm yr $^{-1}$), (c) nature of fan deposits (debris-flow vs. stream-flow). Inferences are made about outlet spacing ($2w$), degree of saturation of basin margin (s), catchment length (L) and average mainstream slope. Calculations can therefore be made of ϕ and κ and compared with modern analogues. In the first example from the Devonian Hornelen Basin, Norway, the basin margin is assumed to be saturated ($s=1$) and the sediment routing system to be closed ($e_f=1$). In the second example from the Pliocene Ridge Basin, California, the influence of basin margin oversaturation ($s>1$) and export of sediment to downfan fluvial systems ($e_f<1$) is examined. In both examples (Table 2), a plausible set of parameter values is sought to match geological observables and modern geomorphological datasets.

The progradation distances of coarse-grained alluvial fans in the superbly exposed Middle Devonian Hornelen Basin (Steel, 1976; Larsen & Steel, 1978; Gloppe & Steel, 1981) can be measured. The Nibbevatnet, Karlskaret and Hjortestegvatnet debris-flow-dominated fans are located along the northern margin of the Hornelen Basin (Fig. 5). The progradation distance of the fan deposits reaches a maximum of ≈ 1.5 km, ≈ 1 km and ≈ 1.5 km, respectively. The fans on the southern

margin are stream-flow dominated. The Borrevatnet, Svartevatnet and Lassenipa fans reach a maximum of 3.5 km, 2.5 km and 2.5 km into the basin, respectively. Fans along the northern margin have a smaller radius (less than half) than those along the southern margin. The time-averaged subsidence rate is thought to have been about 2.5 mm yr $^{-1}$ (Nilsen & McCauglin, 1985).

If the debris-flow fans were perfectly semicircular and were distributed along the mountain front so that they just mutually touched, their spacing can be estimated as 2 km ($w=1$ km). It is necessary to iterate a plausible set of parameter values by using a range of catchment aspect ratios (Talling *et al.*, 1997), invoking Hack's law or later derivatives (Eq. 15), and an average catchment slope between 0.2 and 0.3 (Seidl & Deitrich, 1992) to promote debris flow erosion. The landslide rate constant κ can be calculated making best guesses of the maximum length scale for landsliding (L_1) and compared with long-term values estimated from present-day mountain belts.

If the debris flow-dominated northern fans are taken to have progradation distances of 1 km, their areas must have been ≈ 1.6 km 2 . The range of aspect ratio given by Talling *et al.* (1997) is 1.4–4.1. Catchment lengths therefore vary from 2.8 km to 8.2 km. Using Eq. (15), catchment areas range from 3.1 km 2 to 28 km 2 . This gives

Table 2. Parameter values used and calculated in the palaeocatchment studies of the Hornelen Basin, Norway, and the Ridge Basin, California.

	Hornelen Basin, Norway		Ridge Basin, California	
	Debris-flow northern fans	Streamflow southern fans	Debris-flow western fans (Violin Breccia)	Streamflow north-eastern fans (Ridge Route Fm.)
<i>Observables</i>				
Progradation distance x (km)	1.0	3.0	1.5	4.0
Fan aggradation rate \dot{y} (mm yr $^{-1}$)	2.5	1.5	3.0	1.5
<i>Assumptions</i>				
Saturation s	1	1	0.75	2
Storage efficiency e_s	1	1	1	1
Fan transport efficiency e_f	1	1	1	0.5
Outlet spacing (km)	2.0	6.0	4.0	4.0
Aspect ratio	1.4–4.1	1.4–4.1	1.4–4.1	1.4–4.1
<i>Inferences</i>				
Fan area A_2 (km 2)	1.6	14	3.5	13
Average catchment slope	0.3	0.2	0.3	0.2
Catchment length L (km)	2.8–8.2	8–25	5.6–16.4	5.6–16.4
Maximum length scale for landsliding L_1 (km)	0.3–0.8	0.5–1.0	0.5–1.0	0.5–1.0
<i>Calculations</i>				
Catchment area A_1 (km 2)	3.1–28	27–273	12.9–115	12.9–115
Range crest elevation (m)	840–2460	1600–5000	1680–4920	1120–3280
ϕ	0.52–0.06	0.52–0.05	0.27–0.03	1.0–0.11
Sediment efflux V (mm yr $^{-1}$)	0.9–0.1	0.52–0.05	0.54–0.06	2.0–0.22
κ ($\times 10^{-5}$ km $^{-1}$ yr $^{-1}$)	3.0–0.13	1.0–0.05	1.6–0.1	4.0–0.22
Most likely range of ϕ	0.2–0.1	0.4–0.2	0.2–0.1	0.4–0.2

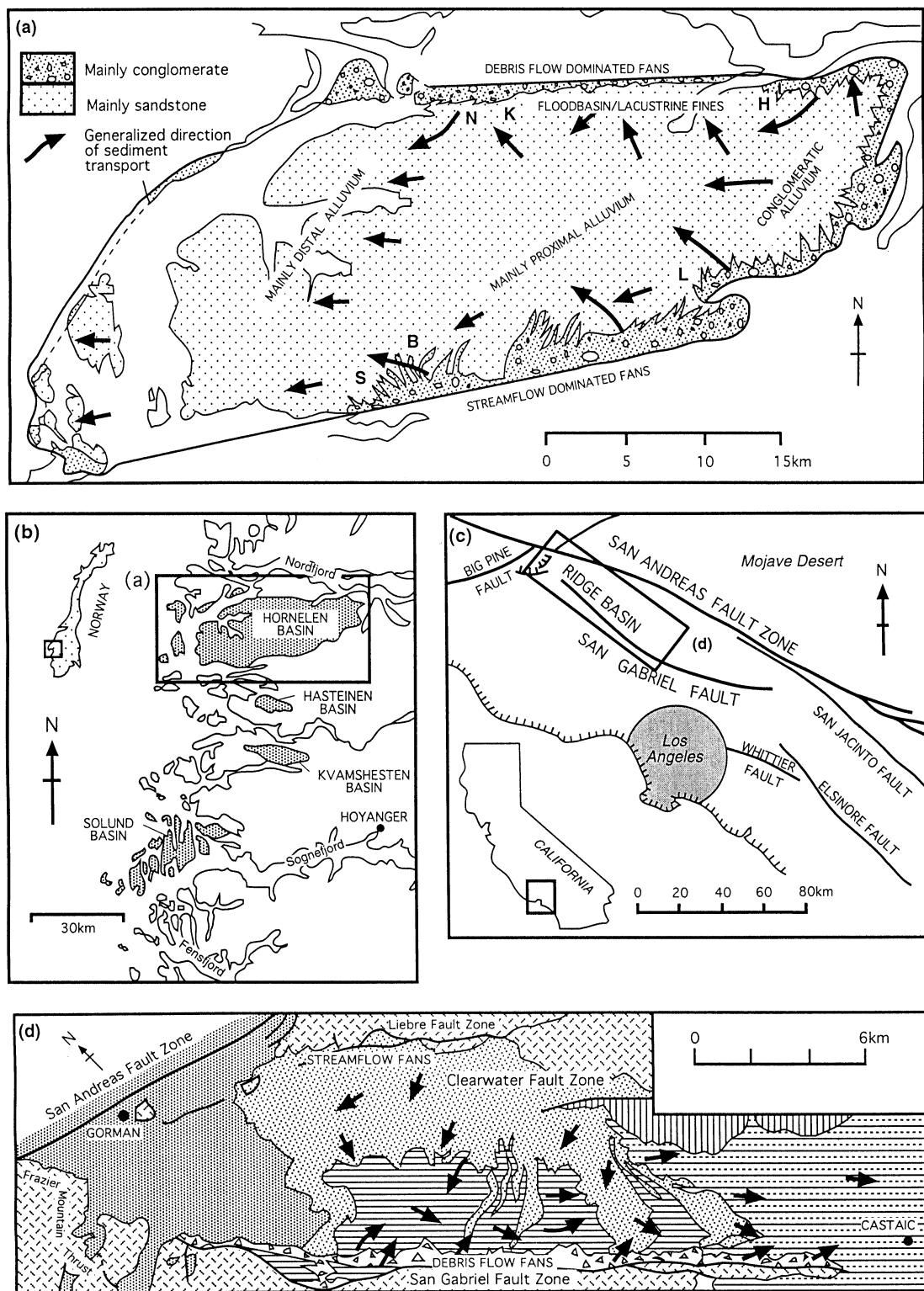


Fig. 5. (a) Devonian Hornelen Basin of western Norway (Gloppen & Steel, 1981) and location in western Norway (b); fans highlighted – B, Borrevatnet; S, Svartevatnet; L, Lassenipa; N, Nibbevatnet; K, Karkskaret; H, Hjortestegvatnet. (d) Mio-Pliocene Ridge Basin of California (Link & Osborne, 1978) and location within San Andreas strike slip system (c).

values of ϕ between 0.52 and 0.06. Assuming the average mainstream slope to be steep enough to generate debris flow erosion (0.3) gives range crest elevations between 840 m and 2460 m. The sediment efflux for the northern catchments where j' is 2.5 mm yr^{-1} ranges between 0.9

and 0.1 mm yr^{-1} . This is the average denudation rate in the catchment. The calculation of κ requires knowledge of the maximum length scale for landsliding. The value depends on rock strength and on the extent of dissection of valley sides. In the catchments bordering the northern

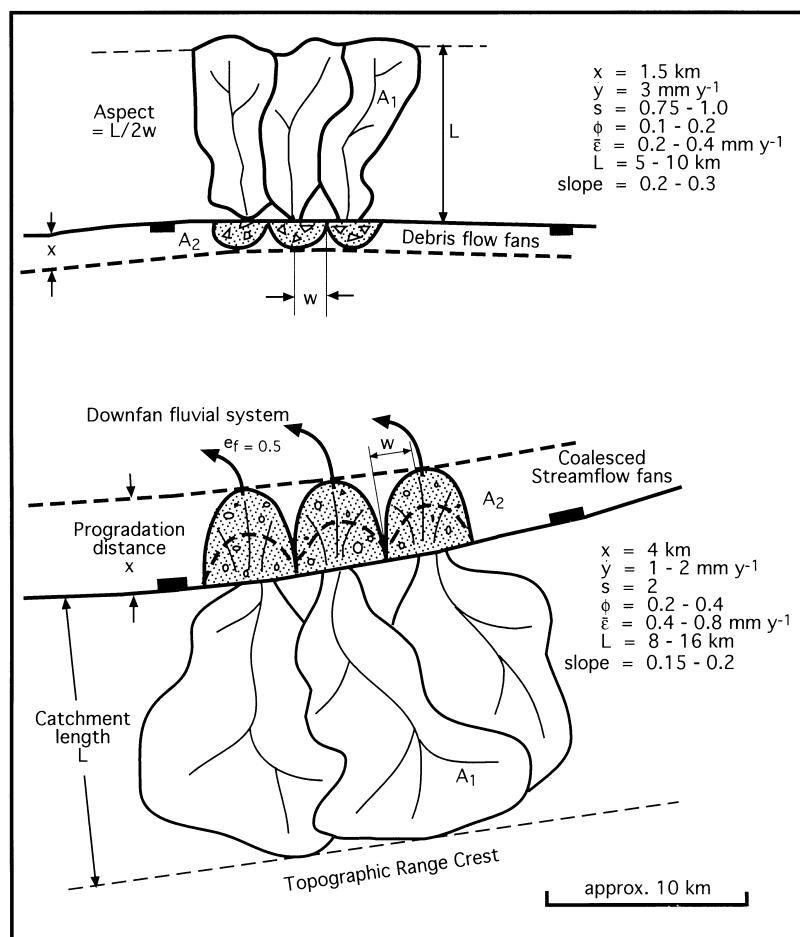


Fig. 6. Schematic illustration of parameter values interpreted for the Ridge Basin case study.

basin edge, we somewhat arbitrarily use $L_1=0.3 \text{ km}$ for the lower end of the range of catchment size and $L_1=0.8 \text{ km}$ for the upper end of the range. This gives a range of κ from 3.0×10^{-5} to $0.1 \times 10^{-5} \text{ km}^{-2} \text{ yr}^{-1}$. The set of parameter values corresponding to the high landslide rate constant of $3 \times 10^{-5} \text{ km}^{-2} \text{ yr}^{-1}$ are unlikely, since they imply excessively high landslide activity comparable to the tectonically extremely active Southern Alps of New Zealand (Hovius *et al.*, 1997). A more plausible set of parameter values is generated with lower values of ϕ between 0.1 and 0.2. Such a set is similar to that of the Black Mountains and their small debris-flow dominated fans along the east side of Death Valley (Denny, 1965). The rectangular shaped catchments (Fig. 4) have average aspect ratios (length/width) in excess of 2.6 and values of ϕ between 0.02 and 0.17. For comparison, the maximum range crest elevation in the Black Mountains is 1946 m (Funeral Peak).

If the progradation distance of the southern streamflow-dominated fans is taken as 3 km, and once again they are assumed to be semicircular in plan, their areas can be estimated as about 14 km^2 . Using the same range of aspect ratio as above, catchment lengths vary from 8 km to 25 km. Using Eq. (15), catchment areas range from 27 km^2 to 273 km^2 , giving a range of ϕ of 0.52–0.05. At an average mainstream slope of at most 0.2 to promote stream flow processes, the maximum range crest

elevations are between 1680 m and 4920 m. The sediment efflux (average denudation rate) for a fan aggradation rate of 1.5 mm yr^{-1} along the southern margin of the basin ranges between 0.5 and 0.05 mm yr^{-1} . Taking the maximum length scale for landsliding to range between 0.5 km for the short catchments and 1 km for the long catchments gives values of κ between 1×10^{-5} and $0.05 \times 10^{-5} \text{ km}^{-2} \text{ yr}^{-1}$. The set of parameter values resulting in range crest elevations of nearly 5 km are not plausible. Consequently, a value of ϕ closer to the upper end of the range appears to be more likely. Catchments in the Panamint Range feeding coalesced streamflow-dominated fans along the west side of Death Valley are pear-shaped with average aspect ratios of about 1.6 and values of ϕ between 0.21 and 0.59. The maximum range crest elevation in the Panamints is 3368 m (Telescope Peak).

Ridge Basin, California (Crowell & Link, 1982), developed in the late Miocene east of a major bend in the transcurrent San Gabriel fault. During the Pleistocene, deformation transferred to the modern trace of the San Andreas fault to the north-east. The Violin Breccia is located along the western margin of the basin and represents steep, debris-flow- and landslide-dominated cones that prograded up to 1.5 km into lacustrine shales (Link & Osborne, 1978). Average accumulation rates were 3 mm yr^{-1} (Nilsen & McCauglin, 1985). The Ridge

Route Formation includes alluvial fans deposited along the north-east flank of the basin adjacent to basement uplifted by the Liebre and Clearwater faults. These fans grade downslope into fluviaatile deposits. The Ridge Basin is therefore similar in many geometrical and stratigraphic characteristics to the Hornelen Basin. Making the same assumptions as for the Hornelen examples would produce quantitative results for catchment sizes, range crest elevations and denudation rates very similar to those of the northern and southern basin margins of the Hornelen. The Ridge Basin can be used, however, to examine the effects of degree of basin margin saturation and fan bypass.

We firstly investigate the effects of undersaturation of the western margin (Violin Breccia) of the Ridge Basin. Parameter values taken from geological observations are $\dot{y}=3 \text{ mm yr}^{-1}$ and $x=1.5 \text{ km}$. The results of using an understauration of $s=0.75$ for the western basin margin are shown in Table 2. The effect of undersaturation is to decrease the calculated values of ϕ together with the sediment efflux (or average denudation rate) required to feed the basin-margin fans. The upper end of the range of values of ϕ (0.27–0.03) requires very high values of κ ; a very low value of ϕ is also unlikely since it necessitates range crest elevations of $\approx 5 \text{ km}$. A preferred range is $0.1 < \phi < 0.2$, with corresponding values of κ from 0.8 to $0.2 \times 10^{-5} \text{ km}^{-2} \text{ yr}^{-1}$ (V from 0.4 to 0.2 mm yr^{-1}).

Second, the impact of fan efficiency and oversaturation along the north-eastern margin can be studied. Let us take geological observables as $x=4 \text{ km}$ and $\dot{y}=3 \text{ mm yr}^{-1}$. Since the alluvial fans grade downslope into fluviaatile deposits, we take a fan efficiency factor of $e_f=0.5$. Assuming that the fans are coalesced along this margin, we also take the saturation to be $s=2$. Over a large part of the calculated range of ϕ (1.0–0.11), the sediment efflux required is unrealistically high. The streamflow-dominated fans on the north-eastern margin of the Ridge Basin can be better modelled if the tectonic subsidence rate is in the region of $1\text{--}2 \text{ mm yr}^{-1}$ (Table 2). Reasonable sediment effluxes (average denudation rates) and landslide rate constants κ can then be produced if ϕ is less than about 0.4. Although oversaturation has an effect on the fan area for a given value of the progradation distance, the main effect of oversaturation is to reduce palaeocatchment sizes, thereby driving up the average denudation rate required to supply basin margin fans. Export of sediment to downfan fluvial systems also increases the efflux required to feed the fan. Consequently, in coalesced (oversaturated) and ‘open’ systems, sediment discharges must be high. Such discharges might result from very high tectonic uplift rates promoting high average denudation rates, or from being gathered from large catchment areas.

In summary, the theoretical model outlined above can be applied to geological contexts where some of the parameter values can be well constrained. A preliminary analysis of the Hornelen and Ridge Basins, both charac-

terized by contrasting styles of alluvial fan on opposing margins of the basin, indicates that the debris-flow-dominated fans were produced in small, steep, relatively narrow catchments feeding saturated or undersaturated, rapidly subsiding basin margins at approximately $0.1 \leq \phi \leq 0.2$. The streamflow-dominated fans on the opposite margin were fed by larger, less steep catchments at $0.2 \leq \phi \leq 0.4$. The lower tectonic subsidence rate at the basin margin and larger sediment discharge to the fan apex allowed fans to coalesce and export sediment to downfan fluvial systems. Average denudation rates modelled are in the region of up to 1 mm yr^{-1} . The values of ϕ estimated from the ancient examples are within the range of values obtained from modern fans, and typify fans in active basin margin settings where tectonic subsidence rates are inferred to be high.

The uncertainties in the palaeocatchment studies above are large and the numerical results should be treated with caution. Despite this, they provide some insights as geological examples where a catchment-fan sediment routing system can be investigated.

CONCLUSIONS

- 1 The sediment discharge from catchment outlets into a basin acts as a boundary condition for far-field sediment dispersal and controls the architecture of basin margin depositional units. The supply control on basin margin geometries can best be studied in the closed or quasiclosed sediment routing systems of uplifting fault block catchments and basin margin fans.
- 2 The sediment output from a mountainous catchment can be estimated from the rate of hillslope erosion. A steady-state landscape can be envisaged where rivers incise into uplifting bedrock, thereby keeping hillslope gradients at the critical value for landsliding. Consequently, the landslide efflux of the catchment is determined primarily by the tectonic uplift rate of rock. A conservation of solid sediment volume allows a model to be developed where the sediment discharge from the catchment is distributed evenly over a fan surface in a basin margin with a uniform tectonic subsidence rate. The model incorporates the effects of sediment storage in the catchment, fan bypass to downslope fluvial systems, and undersaturation or oversaturation of the basin margin.
- 3 A plot of catchment area vs. fan area for a large number of fans in different climatic and tectonic settings shows a very wide scatter of data. The causes of the scatter are investigated through a dimensionless sediment dispersal parameter ϕ which incorporates the effects of tectonic subsidence rate, sediment efflux, and catchment storage and fan bypass efficiencies. A plot of over 100 modern fans in the semiarid to arid SW USA shows ϕ to vary over about two orders of magnitude from 0.1 to 10, and allows the impact of the tectonic subsidence rate to be demonstrated.
- 4 Application to the ancient geological record is fruitful

if some parameter values can be constrained, such as fan progradation distance and time-averaged tectonic subsidence rate. Although it is not possible to produce one unique solution set of parameter values, values can be combined iteratively to reconstruct palaeocatchment sizes and tectonic uplift rates as well as shed light on the possibilities of lateral facies continuity and export to downfan facies tracts. A simple analysis of the Hornelen (Devonian, Norway) and Ridge (Miocene–Pliocene, California) Basins suggests that both basins were characterized by opposing margins with markedly different catchment sizes and transport mechanics, basin subsidence rates, and fan geometries. These differences are encapsulated in the values of ϕ , which vary from 0.1 to 0.2 for the small, steep catchments feeding debris-flow- and landslide-dominated fans along the rapidly subsiding basin edge, and 0.2–0.4 for the larger and shallower catchments supplying sediment to streamflow-dominated fans and downfan fluvial systems along the less rapidly subsiding margin.

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