

Chapter 12

Dryland Rivers: Processes and Forms

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Introduction

Dryland alluvial rivers vary considerably in character. In terms of processes, high energy, sediment-laden flash floods in upland rivers contrast dramatically with the low sediment loads and languid flows of their lowland counterparts while from a form perspective, the unstable wide, shallow and sandy braid plains of piedmont rivers are quite different from the relatively stable, narrow, deep and muddy channels of anastomosing systems (Nanson et al. 2002). It is also apparent that few, if any, morphological features are unique to dryland rivers. The variety of dryland river forms and the absence of a set of defining dryland river characteristics makes it difficult to generalise about dryland rivers and raises questions about whether it is necessary (or even desirable) to consider dryland river systems separately from those in other climatic zones. Indeed, as noted in the introduction to this volume, the recent shift away from the study of morphogenesis within specific climatic regimes (e.g. Tricart and Cailleux 1972) towards the study of geomorphological processes *per se* (e.g. Bates et al. 2005) has largely undermined the distinctiveness of desert geomorphology. This is not to say rivers draining different climatic regions do not differ in aspects of their behaviour. They clearly do, as exemplified in several reviews of tropical (Gupta 1995), periglacial (McEwen and Matthews 1998) and dryland (Graf 1988; Knighton and Nanson 1997; Reid and Frostick 1997; Tooth 2000a) fluvial geomorphology. However, given the diversity of dryland river mor-

phologies, and the fact that many of the forms are shared by rivers that drain other climatic zones, it is far from clear how far dryland rivers can be categorised as a distinctive group and whether such a categorisation provides a suitable basis for developing an understanding of them. On this basis, rather than attempt to understand dryland rivers as a distinctive and definable group of rivers, this chapter seeks explanations for the character (the diversity, distinctiveness and, in some cases, the uniqueness) of dryland rivers in terms of the operation of geomorphological processes as they are mediated by the climatic regime. Because the multivariate and indeterminate nature of river channel adjustment makes it difficult to describe directly the three-dimensional subtleties of channel form, the chapter follows the approach of Ferguson (1981) and concentrates on three separate two-dimensional views in turn: the channel cross-section (size and shape), planform geometry and longitudinal profile. Adjustments to the configuration of channel bed sediments are also considered. Since the form of alluvial rivers evolves in response to the movement of bed material, the chapter starts by considering the dynamics of solute/sediment transport in dryland rivers. A discussion of dryland river hydrology can be found in the preceding chapter.

Solute and Sediment Transport

Although the geomorphological effectiveness of fluvial activity in dryland environments is widely recognised (e.g. Graf 1988; Bull and Kirkby 2002), our understanding of key processes is far from complete. Relatively little is known about the transport of solutes in dryland streams. The paucity of information

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on water chemistry reflects the limited importance of solution to dryland denudational processes (Meybeck 1976) and the fact that solute transport has little effect on channel form and stability. Recent work, however, has highlighted a growing awareness of the importance of floodwater chemistry for aquatic ecology (Grimm et al. 1981; Davies et al. 1994; Costelloe et al. 2005; Smolders et al. 2004), the cycling of nutrients (Sheldon and Thoms 2006) and the properties of alluvial soils (Jacobson et al. 2000a) and geochemical sediments (McCarthy and Ellery 1995; Khadkikar et al. 2000; Nash and McLaren 2003) in dryland environments. Much more is known about the transport of sediments. In terms of the movement of bed material, a basic distinction can be made between sand- and gravel-bed rivers (Parker 2008, p. 178–182). In general, sand-bed rivers are dominated by high excess shear stresses and suspended sediment transport while gravel-bed rivers are dominated by low excess shear stresses and bedload transport. Most work in dryland streams has focussed on understanding the dynamics of suspended sediment transport. Direct measurements of bedload are notoriously difficult to make and the logistical and practical difficulties are enhanced considerably by the uncertainty, unpredictability and infrequency of rainfall and runoff in dryland environments. Consequently there are few data sets documenting the dynamics of bedload transport in dryland rivers.

Dissolved Load

Since dissolved materials mix readily in turbulent flow, solute concentrations in streams and rivers are conventionally determined from a single, mid-stream sample. Solute monitoring programmes usually adopt discrete sampling methods (either manual or automatic) although an increasing number of water-quality parameters can be measured continuously *in situ*. Concentrations of total dissolved solids, for example, are routinely derived from measurements of electrical conductivity (specific conductance SC ; $\mu\text{S cm}^{-1}$) which are converted to ionic concentrations with the aid of ion-specific calibration curves. Since the relationship between electrical conductivity and ionic concentration is temperature dependent, conductance values are usually adjusted to a standard temperature

of 25°C. Concentrations are expressed in units of mass per volume (mg l^{-1}) or mass per mass (parts per million; ppm). The units are usually used interchangeably even though a density correction should be applied to account for variations in fluid density due to temperature and solute concentration (USGS 1993). The unit of micro moles per litre ($\mu\text{mol l}^{-1}$) is used for chemical mass-balance calculations.

The relationship between solute concentration (C_c ; mg l^{-1}) and discharge (Q ; $\text{m}^3 \text{s}^{-1}$) is usually modelled as the power function

$$C_c = aQ^b \quad (12.1)$$

in which the empirical coefficients a and b are fitted by ordinary least squares regression. Typically, $b < 0$ indicating that solute concentrations decrease with discharge (Walling and Webb 1983) reflecting the dilution effect by stormflow of low ionic status. Classic dilution effects have been observed in dryland settings. Hem (1985) for example attributed a decline in electrical conductance in the San Francisco River, Arizona to stormflow dilution of heavily mineralised perennial spring waters (Fig. 12.1a). A mixing model utilising mass balance equations for three sources of runoff (spring water, baseflow and storm runoff) provides a good fit to the observed data. Dilution concepts may, however, be less useful in ephemeral streams due to the absence of base flow and the high velocities of overland flow which limit the length of time runoff has to react to near-surface rock and soil minerals. In the ephemeral flowing Nahal Eshtemoa in Southern Israel, for example, marked variations in solute concentrations are only observed during the rising stages of the flood pulse and are attributed to the flushing of solutes from the watershed at the onset of the event (Fig. 12.1b). Thereafter, the ionic concentrations and composition of rainfall and runoff are broadly comparable. Similar flushing effects have been observed in the Gila River near Fort Thomas in Arizona (Hem 1948) and may explain the chemical changes observed during flash floods in Sycamore Creek (Fig. 12.1c) and KR Wash, also in Arizona (Fisher and Minckley 1978; Fisher and Grimm 1985). The data from Sycamore Creek also highlights the contrasting behaviour of different solutes during flood events.

In a spatial context, solute concentrations in dryland rivers have been found to increase downstream due to flow attenuation by transmission losses (Jacobson

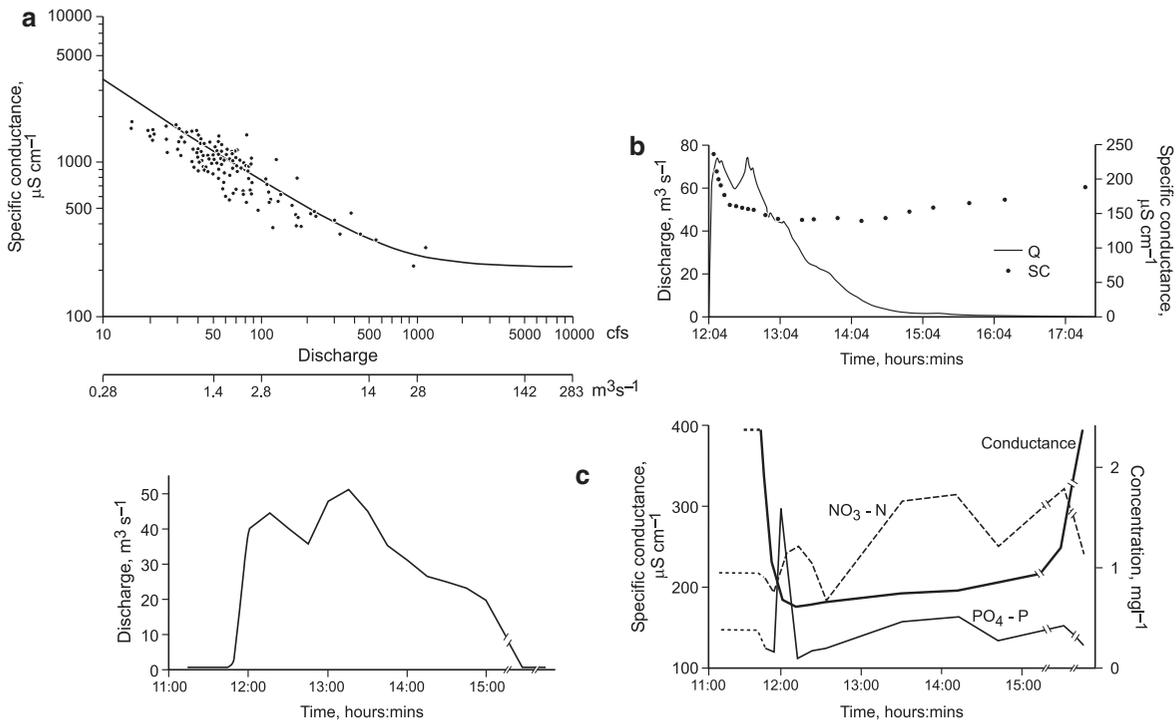


Fig. 12.1 Chemical characteristics of flood waters in (a) San Francisco River, Arizona (after Hem 1985), (b) Nahal Eshtemoa, Israel (after Alexandrov 2005) and (c) Sycamore Creek, Arizona

(after Fisher and Minckley 1978). The curve in (a) represents the fit of a three-source mixing model to the data

et al. 2000b). In terms of organics, several authors have noted that dryland rivers transport high concentrations of dissolved and particulate organic matter (Jones et al. 1997; Jacobson et al. 2000b). The comparatively high organic matter loadings of dryland streams have been attributed to lower mineralisation rates, limited sorption of dissolved organic matter in sandy soils and the rapid concentration of runoff into channels (Mullolland 1997).

Suspended Sediment

The properties of suspended sediment are usually measured using extracts of water-sediment mixtures (Edwards and Glysson 1999) but they can also be measured *in situ* using optical sensors (Gippel 1995). There are significant spatial and temporal heterogeneities in suspended sediment concentrations so measurements must be made within an appropriate sampling frame-

work to prevent sampling bias (Meade et al. 1990; Hicks and Gomez 2003).

Although suspended sediment loads reflect a wide range of climatic, topographic, lithological and anthropogenic controls (Lvovich et al. 1991), rivers draining areas of low precipitation are frequently distinctive in terms of high suspended sediment concentrations (Walling and Kleo 1979; Alexandrov et al. 2003). One of the most sediment-laden rivers on Earth is Rio Puerco in semi-arid New Mexico. Concentrations in excess of 600,000 ppm are routinely measured at the USGS gauging station near Bernardo and the 50-year average annual suspended sediment concentration of 113,000 ppm ranks fourth highest in a global comparison of sediment load data for selected world rivers (Gellis et al. 2004). In fact, the transport of hyperconcentrations of suspended sediment (defined as those in excess of 400,000 ppm) is a frequent occurrence in many dryland rivers (Lane 1940; Beverage and Culbertson 1964; Gerson 1977; Stoddart 1978; Walling 1981; Lekach and Schick 1982; Xu 1999).

As for solutes, the relation between concurrent measurements of suspended sediment concentration (C_s ; mg l^{-1} or ppm) and discharge is conventionally modelled by a power function (Equation 12.1). Unlike the solute case, however, the exponent (b) is typically greater than zero indicating that suspended sediment concentrations increase with discharge (Fig. 12.2).

Frostick et al. (1983) note that suspended sediment rating curves for dryland streams are associated with higher coefficients (a) and lower exponents (b) than those derived for humid-temperate streams. The difference in coefficients is indicative of the transport of larger suspended sediment loads at low discharges. The difference in exponents indicates that dryland suspended sediment concentrations are less sensitive to changes in discharge. In fact, sediment concentrations in dryland environments increase at a rate less than a proportionate increase in discharge ($b < 1$) which is in contrast to humid-temperate streams for which b normally lies in the range 2–3 (Leopold and Maddock 1953).

Measurements of suspended sediment incorporate both wash load and suspended bed material load. The former is fine-grained sediment, typically fine sands, silts and clays, delivered to the channel with hillslope runoff. The latter comprises coarser material sourced from the instream sediments. Although the two components cannot be separated unequivocally, an arbitrary distinction can often be made on the basis of sediment size by comparing bed material and suspended sediment grain size distributions (Fig. 12.3).

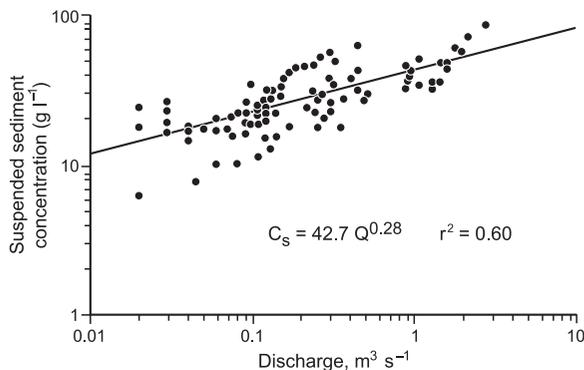


Fig. 12.2 Suspended sediment rating curve for the Katorion catchment, central Kenya (after Sutherland and Bryan 1990). r^2 is the coefficient of determination of the fitted rating relation

Suspended Bed Material Load

It has long been appreciated that bed material is suspended into the flow by the action of coherent turbulent flow structures or eddies (e.g. Sutherland 1967; Jackson 1976). More recent work has illuminated the hydrodynamics of particle suspension in considerable detail. Over smooth boundaries (e.g. planar beds of sand-sized sediment), eddies originate as hairpin vortices from alternate zones of high and low speed within the viscous sublayer (see review by Smith 1996). Similar structures are observed in flows over gravel-sized sediment and over bedforms due to the shedding of wakes from individual clasts (Kirkbride 1993) and from the shear layer that forms due to flow separation downstream of bedform crests (McLean et al. 1996). Of particular importance to the suspension of bed material is the violent ejection of low momentum fluid from the bed during turbulent motions (Lapointe 1992; Garcia et al. 1996). These flow ejection events, or ‘bursts’, lift particles into the flow and oppose the tendency for the uplifted grains to settle under the influence of gravity.

According to this model, a particle will remain suspended in the flow providing the vertical turbulent velocity fluctuations exceed the particle’s fall velocity. Consequently, the competence of a turbulent flow to transport sediment in suspension is commonly defined by the criterion

$$v'/\omega_0 > 1 \quad (12.2)$$

in which v' is the maximum root-mean-square vertical turbulent velocity fluctuation (m s^{-1}) and ω_0 is the mean settling velocity of the suspended sediment (m s^{-1}). For shear turbulence, there is abundant experimental evidence that the upward components of vertical velocity fluctuations (v_{up}' ; m s^{-1}) are, on average, greater than the downward components (v_{dn}' ; m s^{-1}) and that v' is proportional to the shear velocity, u_* (m s^{-1} ; McQuivey and Richardson 1969; Kreplin and Eckelmann 1979). Using the assumptions $v_{up}' = 1.6v'$ and $v' = 0.8u_*$, Bagnold (1966) expressed Equation 12.2 in terms of Shields’ dimensionless bed shear stress, τ^* ,

$$\tau^* = \frac{\tau}{(\rho_s - \rho_f)gD} \quad (12.3)$$

where τ = bed shear stress (N m^{-2}), ρ_s = density of sediment (kg m^{-3}), ρ_f = density of flow (kg m^{-3}),

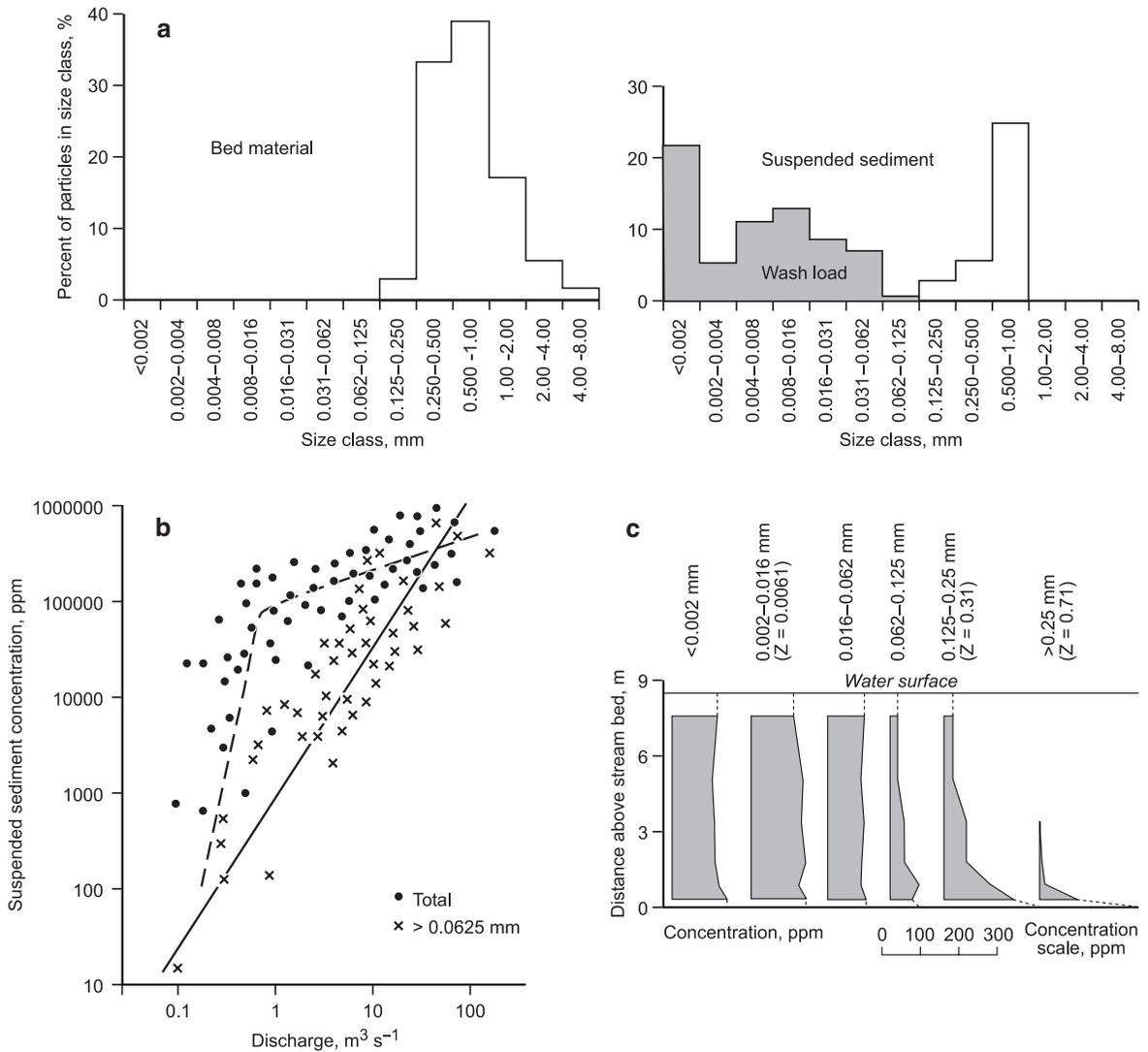


Fig. 12.3 Using grain size to distinguish between wash material and bed material in suspended sediments. (a) Bed material and suspended sediment size distributions in the Rio Grande at Otowi Bridge, New Mexico (after Nordin and Beverage 1965). Suspended sediment finer than 0.125 is not represented in the bed and is assumed to be wash load. (b) Wash load and suspended sediment concentrations in the Paria River at Lees Ferry, Arizona 1954–1965 (after Gregory and Walling 1973). Wash load

concentrations frequently depart from simple bivariate rating relations (Equation 12.1) because of catchment controlled variations in sediment supply. (c) Vertical distribution of different sediment sizes in the Mississippi River at St Louis, Montana (after Colby 1963). Vertical concentration gradients are commonly uniform for wash load and steeply decreasing away from the bed for suspended bed material load. Values of the Rouse number (Z) from Allen (1997 p. 197)

g = acceleration due to gravity ($m s^{-2}$) and D = particle size (m) to give a suspension threshold, τ_s^* :

$$\tau_s^* = \frac{\omega_o^2}{1.56 (\rho_s/\rho_f - 1)gD} \quad (12.4)$$

For grain and flow densities of 2750 and $1000 kg m^{-3}$ respectively, the expression simplifies to

$$\tau_s^* = \frac{0.4 \omega_o^2}{gD} \quad (12.5)$$

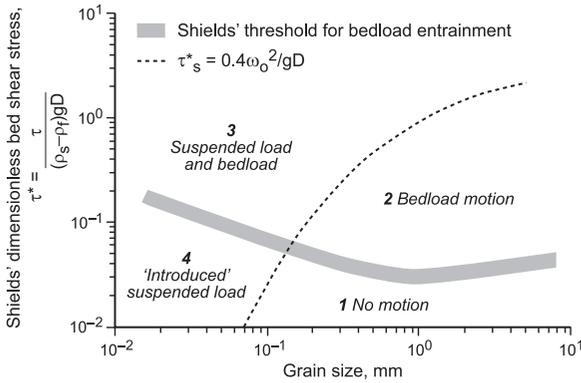


Fig. 12.4 Plot of the Bagnold (1966) suspension criterion along with the Shields bedload entrainment function (after Leeder et al. 2005). Alternative suspension criteria are reviewed by Garcia and Parker (1991)

The Bagnold suspension criterion (Equation 12.5) is plotted in Fig. 12.4 along with the Shields curve for the initiation of bedload movement as defined by Miller et al. (1977). Taken together, the two competence criteria define four fields of sediment transport. Fields 1–3 relate to the transport of bed material. Field 4 relates to the transport of material already held in suspension (i.e. the wash load). Since the transport of wash load is dependent on rates of sediment supply from catchment hillslopes rather than the competence of the flow, a functional understanding of the suspended bed material load in relation to channel hydraulics is restricted to field 3 (Vetter 1937; Einstein and Chien 1953).

Methods for predicting the transport rate of suspended bed material from flow and sediment characteristics are based on models for the vertical concentration and velocity profile in steady uniform flow (e.g. Einstein 1950; van Rijn 1984; Fig. 12.5).

The suspended sediment transport rate per unit width (q_s) is given by

$$q_s = \int_{y'}^Y u C_s \delta y \tag{12.6}$$

where u = flow velocity ($m\ s^{-1}$); y' is a near-bed reference height (m), Y = flow depth (m) and y = height above the bed (m). The vertical velocity profile for a steady uniform flow is given by

$$\frac{u_y}{u_*} = \frac{1}{\kappa} \log e \left(\frac{y}{y_o} \right) \tag{12.7}$$

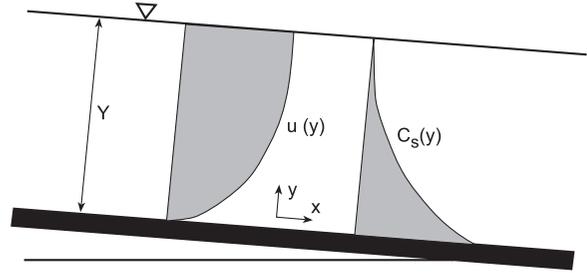


Fig. 12.5 Sketch of velocity (u ; $m\ s^{-1}$) and sediment concentration profiles in steady, uniform open-channel flows (after Wright and Parker 2004a). The vertical profiles of velocity and sediment concentration can be modelled by the law of the wall (Equation 12.7) and the Rouse equation (Equation 12.8) respectively. Y and y are flow depth (m) and height above bed (m) respectively and x is distance downstream (m)

where u_y = velocity at height y ($m\ s^{-1}$), u_* = shear velocity ($m\ s^{-1}$), κ is Von Karman’s constant (≈ 0.4) and y_o is the roughness length (m). Notwithstanding recent advances in the dynamics of sediment suspensions outlined above, models for the vertical suspended sediment concentration profile rely on classical diffusion theory. The Rouse equation balances the downward settling of grains under gravity with their upward diffusion due to turbulence to yield (Rouse 1937):

$$\frac{C_{s_y}}{C_{s_{y'}}} = \left[\frac{(Y - y)(Y - y')}{y y'} \right]^Z \tag{12.8}$$

where C_{s_y} is the concentration at height y ($mg\ l^{-1}$), $C_{s_{y'}}$ is the concentration at height y' ($mg\ l^{-1}$) and Z is a dimensionless suspension parameter known as the Rouse number. Since

$$Z = \omega_o / \beta \kappa u_* \tag{12.9}$$

where β is the sediment diffusion coefficient (commonly assumed ≈ 1), the Rouse number models the concentration gradient by expressing the interaction between the upward-acting turbulent forces and the downward-acting gravitational forces. As shown in Fig. 12.3c, low values of Z model the near-uniform concentrations that result from fine particles (low ω_o) and high flow intensities (high u_*). Conversely, higher values of Z model the stronger concentration gradients generated by larger particles and lower flow intensities.

Application of Equation 12.8 requires definition of $C_{s,y'}$, a reference concentration at height y' above the bed. In the formulation of Einstein (1950), $C_{s,y'}$ is defined at a distance $y' = 2D$ using the Einstein bed-load equation. This, however, has been shown to over-predict near-bed concentrations (Samaga et al. 1986). A number of alternative entrainment functions have subsequently been developed. Of these, Garcia and Parker (1991) conclude that the functions of Smith and McLean (1977) and van Rijn (1984), together with their newly developed relation, performed best when tested against a standard set of data.

Significant improvements to the models for u_y and $C_{s,y}$ (Equations 12.7 and 12.8) have resulted from a consideration of density stratification and bedform effects (e.g. McLean 1991; 1992; Wright and Parker 2004a,b). Sediment-induced density gradients dampen turbulence and reduce the flux of mass and momentum within the water column. The result is an increase in mean flow velocity and a decrease in mean sediment concentration. Since the concentration effect dominates, the net effect is a reduction in transport rate. Stratification also results in finer distributions of suspended sediment because the largest sizes have the strongest concentration gradients and are affected the most by the reduction in vertical mixing. Bedform effects reflect the hydraulic consequences of form drag. For a given mean flow velocity, form drag increases

the total drag (and hence the carrying capacity of the flow) while decreasing skin friction (the ability of the flow to entrain sediment). Because the former usually dominates, the net effect is also a reduction in transport rate. In the suspended load equation of Wright and Parker (2004b), density stratification effects in both velocity and concentration profiles and the effects of bedforms on flow resistance are addressed using relations based on Wright and Parker (2004a) and Engelund and Hansen (1967) respectively. Estimates of near-bed sediment concentrations are made using a modified version of the entrainment function presented in Garcia and Parker (1991). The relation was tested using the data of Toffaleti (1968). Although the model yields reasonably good predictions of suspended sediment concentrations and size distributions, the test is restricted to relatively low concentrations ($C_s < 600 \text{ mg l}^{-1}$). It remains to be seen whether the equation can successfully predict the higher suspended sediment concentrations commonly found in dryland fluvial systems. One model that has been tested in a dryland environment is that due to Laursen (1958). The model was tested by Frostick et al. (1983) in the Il Kimere, a sand-bed stream in the semi-arid province of northern Kenya. The semi-empirical relation makes good predictions of suspended sediment concentrations for those size classes that make up the bed material (Fig. 12.6a). Applications of the

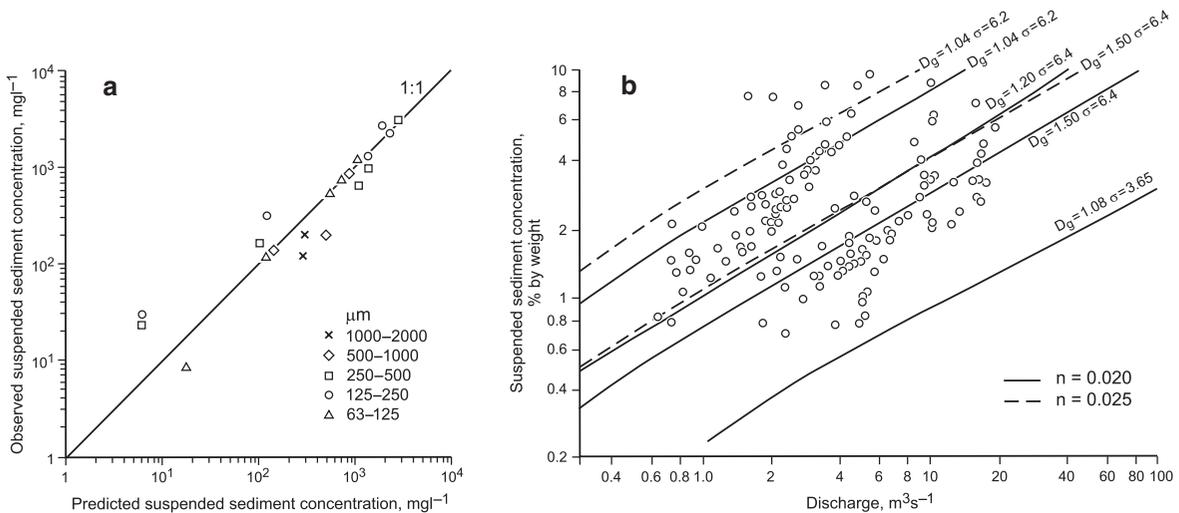


Fig. 12.6 Relations between observed and predicted concentrations of suspended bed material in (a) Il Kimere Kenya (after Frostick et al. 1983) and (b) Walnut Gulch, Arizona (after Renard and Laursen 1975). Predictions due to Laursen (1958). Fig-

ure 12.6b shows the sensitivity of the predicted concentrations to variations in bed material size (as modelled by the mean (D_g) and standard deviation (σ) of the size distribution) and channel roughness (as modelled by Manning's roughness coefficient; n)

model in Walnut Gulch in SE Arizona were also reasonable, (Fig. 12.6b) though the model is very sensitive to poorly constrained input parameters such as bed material size distribution and channel roughness.

It is worth noting that several sediment- and hydraulic-related phenomena of particular relevance to dryland fluvial systems have yet to be incorporated into models of suspended sediment transport. For example, dryland flow events are often unsteady and commence as a flood bore travelling over a dry bed. Recent research into the turbulence characteristics of unsteady flows has demonstrated that turbulence is higher on the rising limb of a hydrograph than it is on the falling limb (Song and Graf 1996; Nezu et al. 1997) with potential consequences for differential suspension of bed material during rising and falling flood stages. Other workers have highlighted the potential for turbulence-induced scouring at the front of advancing bores (Capart and Young 1998). This may explain the finding that peak suspended sediment concentrations in floods that propagate over a dry bed are often associated with the bore rather than the peak discharge (e.g. Frostick et al. 1983; Dunkerley and Brown 1999; Jacobson et al. 2000b). If so, it suggests that the increase in turbulence at the bore is more than sufficient to counteract any reduction in transport capacity due to the entrainment of air into the bore and the consequent reduction in relative sediment density (Chanson 2004). Finally, Dunkerley and Brown (1999) speculate that the infiltration of sediment suspensions into unsaturated porous bed material may be an important mechanism controlling suspended sediment concentrations in dryland rivers. Confirmation of this phenomenon and elucidation of the controls requires careful study of the infiltration of sediment suspensions into unsaturated porous bed materials.

Wash Load

Since a considerable proportion of the sediment carried in suspension by dryland rivers is fine-grained wash load, the suspended sediment dynamics of many dryland streams are complicated by issues pertaining to the availability of sediment on hillslopes. Figure 12.7, for example, shows the variation in

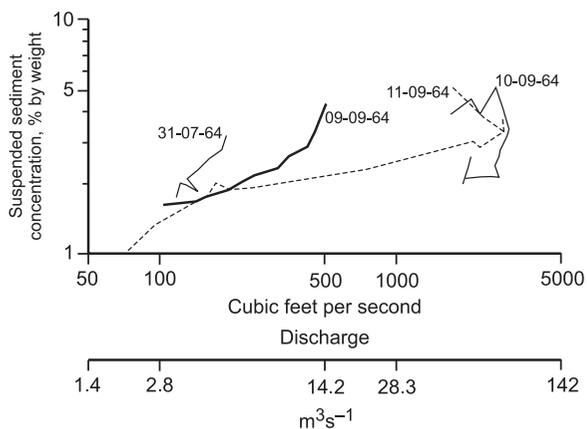


Fig. 12.7 Variations in suspended sediment concentrations during four storms in Walnut Gulch, Arizona (after Renard and Laursen 1975). Compare with Fig. 12.6b

suspended sediment concentrations with discharge during four storm events in Walnut Gulch in south eastern Arizona. It is apparent that concentrations vary by almost an order of magnitude at any specified discharge. Since suspended sediment concentrations for individual events are higher during rising stages than they are at similar discharges during falling stages, much of the scatter can be attributed to clockwise hysteresis in flood-period suspended sediment transport.

Similar storm-period variations in dryland suspended sediment concentrations have been observed in Upper Los Alamos Canyon, New Mexico (Malmon et al. 2004), several central Kenyan rivers (Syrén 1990; Sutherland and Bryan 1990; Ondieki 1995), the Nahal Eshtemoa, Israel (Alexandrov et al. 2003; 2006), Sycamore Creek, Arizona (Fisher and Minckley 1978) and the Burdekin River, Queensland (Amos et al. 2004) and have been attributed to the flushing and subsequent depletion of the most readily mobilised sediment following the generation of runoff on hillslopes and in channels. Sediment supply issues are also important at longer-time scales. Khan (1993) attributed the seasonal decline in suspended sediment concentrations in the Sukri and Guhiya Rivers in western Rajasthan, India to the progressive exhaustion of fine grained sediment deposited on catchment hillslopes by aeolian processes during the preceding dry season (see also Amos et al. 2004). Time-conditioned processes of sediment accumulation and subsequent depletion have also been shown to control suspended

sediment concentrations in the hyper-arid Nahal Yael in southern Israel (Lekach and Schick 1982). In this case, much of the sediment load is sourced from the products of hillslope weathering rather than aeolian deposition.

In many of these studies, the inter- and intra-event time-dependencies in suspended sediment concentrations are strongest for, or even exclusive to, the fine fractions sourced from outside the channel (i.e. the wash load). In other dryland streams, however, hydrologic control has been shown to extend across the full range of grain sizes so that the behaviour of wash and suspended bed material is not so different. In Il Kimere, for example, suspended sediment concentrations for individual size classes, including wash material, show good correlations when rated against discharge (Frostick et al. 1983; Fig. 12.8a). Since coarser fractions are associated with progressively steeper rating relations, the suspended sediment size distribution changes systematically with the flow (Reid and Frostick 1987; Fig. 12.8b).

The hydraulic control of overall suspended sediment concentrations and grain size in Il Kimere is attributed to the combined influence of abundant and readily transportable sediment of all sizes on sparsely vegetated hillslopes and in unarmoured sandy channel fills and the efficiency and effectiveness by which overland flow routes sediment into the channel network. Similar factors were invoked by Belperio (1979) to explain the high correlation observed between wash load concentration and discharge in the Burdekin River, Australia.

Notwithstanding these studies, the general implication of the work discussed above is that catchment-controlled sediment supply issues are significant controls on suspended sediment behaviour in dryland environments. As illustrated by Alexandrov et al. (2003; 2006) an improved understanding of suspended sediment dynamics in dryland streams requires the development of supply-based models that account for the distribution of sediment sources and the spatio-temporal complexity of rainfall-runoff patterns within

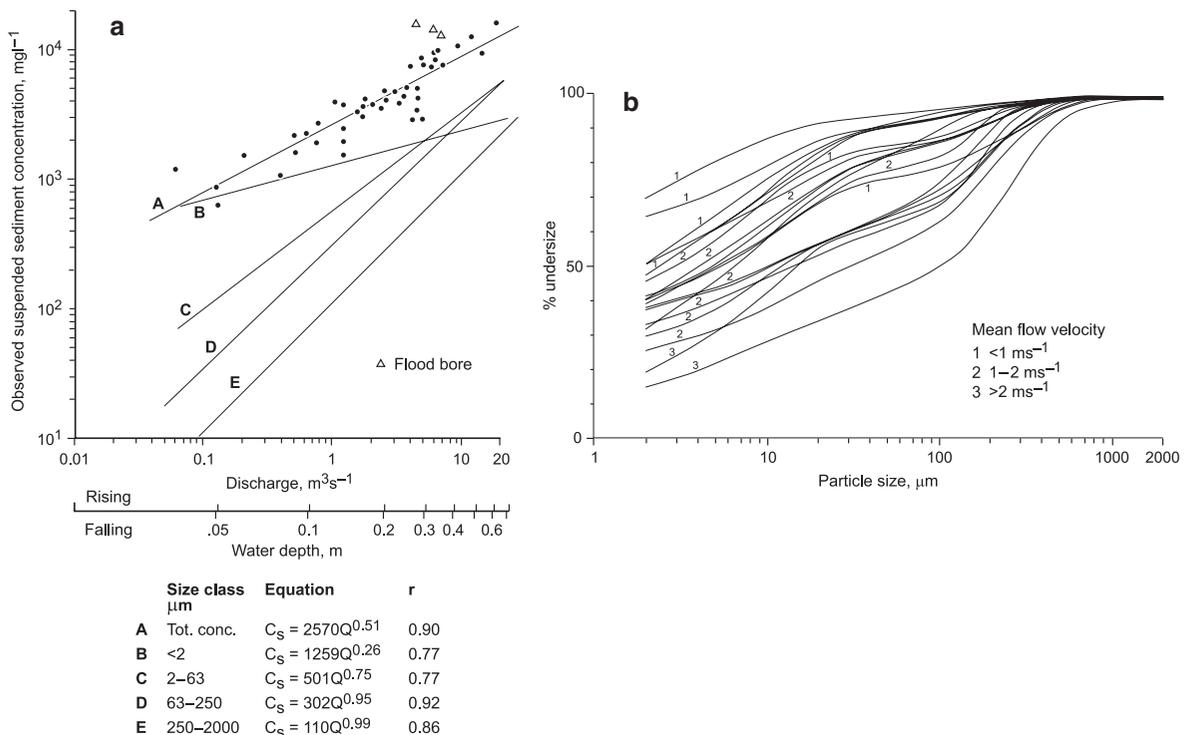


Fig. 12.8 Relation between (a) flow and suspended sediment concentration by size class (after Frostick et al. 1983) and (b) flow and suspended sediment size distribution (after Reid and Frostick 1987) for the Il Kimere Kenya. The curves in b are labelled according to contemporary mean flow velocities. The

progressive shift and change in shape with decreasing flow velocity reflects the dropping-out of coarse bed-material entrained by turbulent suspension at peak flows and the increasing dominance of finer sediment generated by wash processes on catchment hillslopes

dryland catchments. Such models should also account for drainage net influences on water and suspended sediment delivery which have been shown to control the sedimentological character of channel fills (Frostick and Reid 1977) and the type of hysteretic pattern exhibited by suspended sediment rating curves (Heidel 1956).

Bedload Transport

A wide variety of indirect and direct approaches have been used to study bedload transport processes in dryland rivers. Indirect approaches, including reservoir sedimentation studies and particle tracing programmes, are useful in that they do not require personnel to be onsite during flow events. This is of considerable advantage given the ephemeral discharge regime of many dryland fluvial systems. They also provide data that integrate hydrologic, hydraulic and sedimentological responses over a wider range of spatial and temporal scales than is usually possible using direct methods. As a consequence, however, much detail relating to the hydrodynamics of bedload transport processes is lost which can compromise understanding (Schick and Lekach 1993). Such information can be gained from direct and contemporaneous measurements of bedload transport rates and hydraulic parameters during flow events. Although this is an onerous and difficult undertaking in environments where floods are infrequent and unpredictable and where access may be restricted, many of the practical and logistical constraints can be overcome by using automated sampling technologies.

Indirect Measurement Methods

Reservoirs are effective sediment traps and conventional terrestrial and/or bathymetric surveys of reservoir sedimentation provide a well-tested methodology for assessing sediment delivery processes and yields in dryland catchments (Laronne 2000; Haregeweyn et al. 2005; Griffiths et al. 2006). Although most studies do not distinguish between sediment delivered as bedload and as suspended load, such a distinction can often be made since the coarser bedload is generally deposited in prograding

deltaic lobes at the reservoir entrance whilst the finer suspended sediment disperses and settles throughout the reservoir. A reservoir survey was used to quantify the bedload yield of Nahal Yael in the hyper arid southern Negev Desert (Schick and Lekach 1993). The volume of sediment stored within the reservoir delta over a 10-year period was equivalent to a bedload yield of $116 \text{ t km}^{-2} \text{ yr}^{-1}$ which represented two-thirds of the total sediment yield for the 0.5 km^2 catchment. Although bedload is commonly believed to be more significant in dryland environments than it is in humid-temperate environments (Schumm 1968), the ratio of bedload to suspended load is generally less than 0.5 (Graf 1988 p. 139; Powell et al. 1996). The dominance of the bedload contribution to Nahal Yael's sediment yield can be attributed to high magnitude events, steep hillslopes and channels, an abundant supply of coarse-grained sediment on debris-mantled hillslopes and in channel bars and strong hillslope-channel coupling.

Particle tracing techniques (Hassan and Erginzinger 2003) can be used with relative ease in ephemeral rivers because the nature of the discharge regime facilitates tracer relocation and recovery after flood events. Most work has focused on the movement of gravel-sized sediment because of the technical difficulties associated with tagging and tracing finer particles. Detailed tracer-studies of bedload movement in gravel-bed dryland streams have been undertaken in the Negev and Judean Deserts of Israel (Hassan 1990, 1993; Hassan et al. 1991). The results indicate that the travel distances of individual particles during individual events are not correlated with particle size (Fig. 12.9a). Although this conclusion is consistent with field studies in humid temperate environments (e.g. Stelczer 1981) and reflects the stochastic nature of sediment transport (Einstein 1937), it should be noted that the narrow tracer distributions rather precludes an examination of the relative mobility of different sizes. In terms of mean travel distances, the data conform to models of size selective bedload transport in which mean travel distances of particles in the i th size class (\bar{L}_i , m) decrease with increasing mean particle size of that class (D_{gi} , m), though significant departures are observed from the simple $\bar{L}_i \propto 1/D_{gi}$ relation that arises from traditional force balance analyses (Fig. 12.9b). In particular, particle travel distances for the finer sizes are relatively insensitive to particle size. This result has been confirmed by

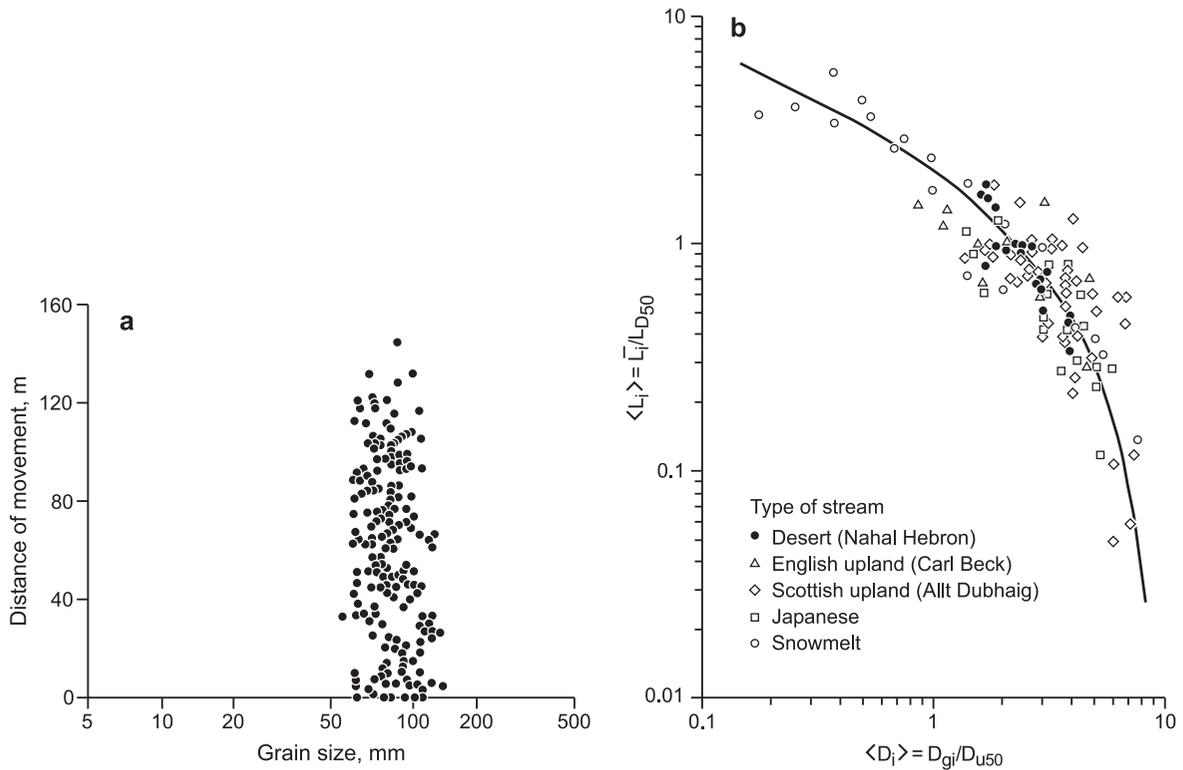


Fig. 12.9 Distance of travel versus particle size (after Hassan and Church 1992). (a) Data for individual clasts, in Nahal Hebron, Israel. (b) Relation between scaled travel distance ($\langle L_i \rangle = \bar{L}_i / \bar{L}_{D_{50}}$) and scaled particle size ($\langle D_i \rangle = D_{gi} / D_{u50}$) for gravel-

bed rivers from a range of hydroclimatic regimes. $\bar{L}_{D_{50}}$ is the mean distance of movement for the particle size group that includes the median size of the surface bed material (D_{50} ; m) and D_{u50} is the median grain size of the subsurface material (m)

Wilcock (1997) and Ferguson and Wathen (1998) and is attributed to the trapping-action of the bed-surface pocket geometry which principally affects the finer sizes (Einstein 1950).

Distributions of particle travel distances were found to conform to the Poisson-based model of Einstein (1937) and Sayre and Hubbell (1965; EHS) and to the two parameter gamma function (Fig. 12.10). The former yielded skew-peaked distributions, whilst the latter gave monotonic (Fig. 12.10a–c) and skew-peaked distributions (Fig. 12.10d–h). The monotonic distributions were associated with relatively small events in which a large number of particles moved only a short distance. The skew-peaked distributions were generated by the larger events in which particle movements were more significant. The skewed models did not fit the data as well as the monotonic models. It is suggested, therefore, that the distributions are only suitable for modelling the local dispersion of sediment. More complex models are required to model

the longer travel distances of large events because of complex bedload-bedform interactions such as the movement of sediment into storage within bars (see also Leopold et al. 1966; Hassan et al. 1999). Particle travel distances are also affected by the sedimentological environment: particles locked within the surface layer, or buried within the subsurface material, travel, in general, shorter distances than unconstrained particles (Hassan 1993).

Tracers are often buried (e.g. Hassan 1990; Hassan and Church 1994) as a result of scour and fill of the stream-bed. Although scour and fill are characteristic of all alluvial rivers, they are of particular significance in many dryland environments where there is often an unlimited supply of sand and fine gravel that is readily entrained by infrequent, but intense flooding (Leopold and Maddock 1953; Colby 1964; Foley 1978). Perhaps the most extensive study of scour and fill in a dryland channel is due to Leopold et al. (1966) who measured stream-bed scour and fill at 51 cross-sections

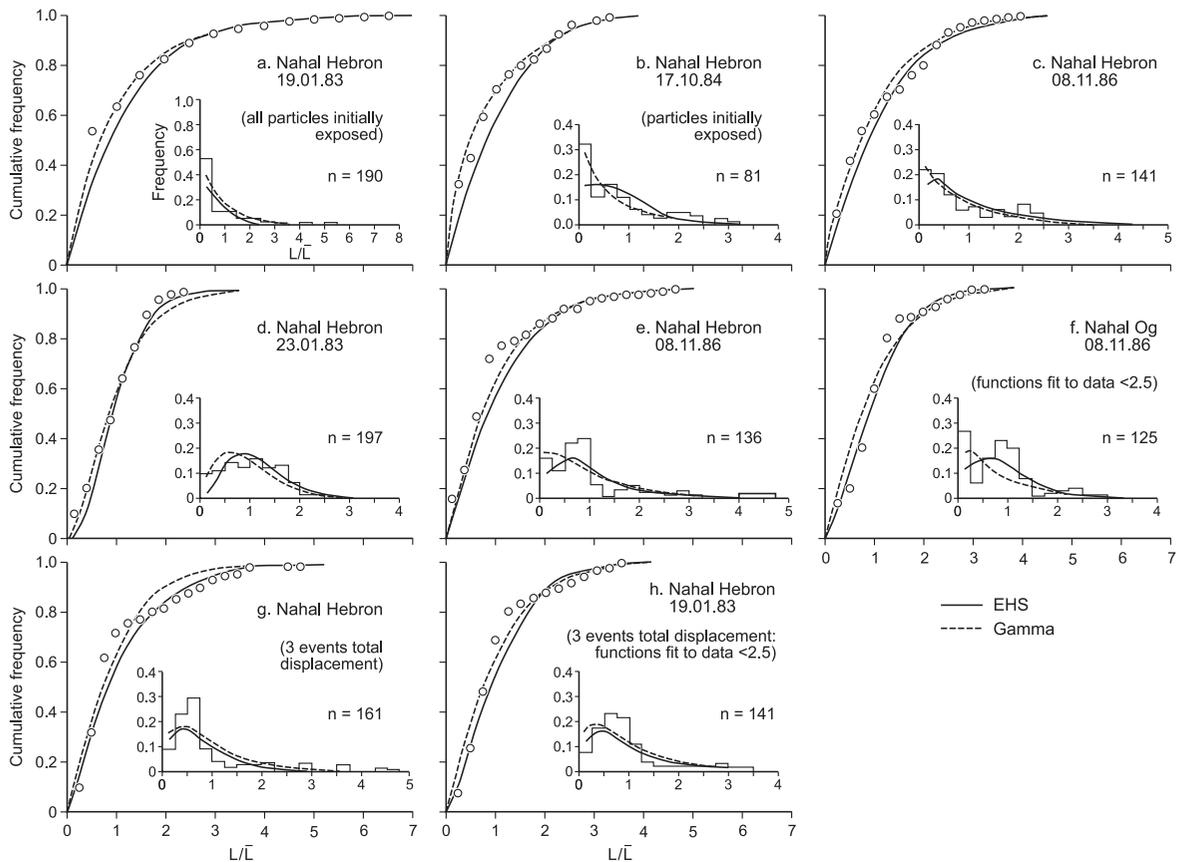


Fig. 12.10 Distributions of particle travel distances in two gravel-bed streams in Israel (after Hassan et al. 1991). Travel distances (L ; m) have been normalised by the mean distance of

movement (\bar{L} ; m). n is the number of data. Only those particles that moved are considered. Similar results are obtained for the full data set

within a 10 mile section of a predominantly sand-bedded arroyo in New Mexico. The results suggest that the bed was scoured extensively during flood events (mean scour depths varied with the square root of discharge per unit channel width) but that compensating fill maintained the channel in approximate balance. More intensive investigations into the variability and pattern of stream-bed scour and fill at channel-reach scales were conducted by Powell et al. (2005, 2006, 2007). These studies deployed dense arrays of scour chains in three low-order channels of the Walnut Gulch catchment in SE Arizona. Detailed statistical analyses demonstrated that mean depths of scour increased with event magnitude and that many populations of scour depths were exponentially distributed (Fig. 12.11a). Exponential model parameters (α ; cm^{-1}) collapse onto a general trend when rated against shear stress in ex-

cess of a threshold shear stress for entrainment (τ_c), thereby providing a means to estimate depths of scour in comparable streams (Fig. 12.11b). In terms of spatial patterns, active bed reworking at particular locations within the reaches resulted in downstream patterns of alternate shallower and deeper area of scour (Fig. 12.11c). During each event, compensating fill returned the streams to preflow elevations indicating that the streams were in approximate steady state over the period of the study (Fig. 12.11d). The results support the suggestion of Butcher and Thornes (1978) that sediment storage does not exert a significant control on sediment transfers through steep headwaters of dryland channels.

Because of the ephemeral discharge regime, the beds of dryland streams are readily accessible and post-event measurements of particle travel distances and

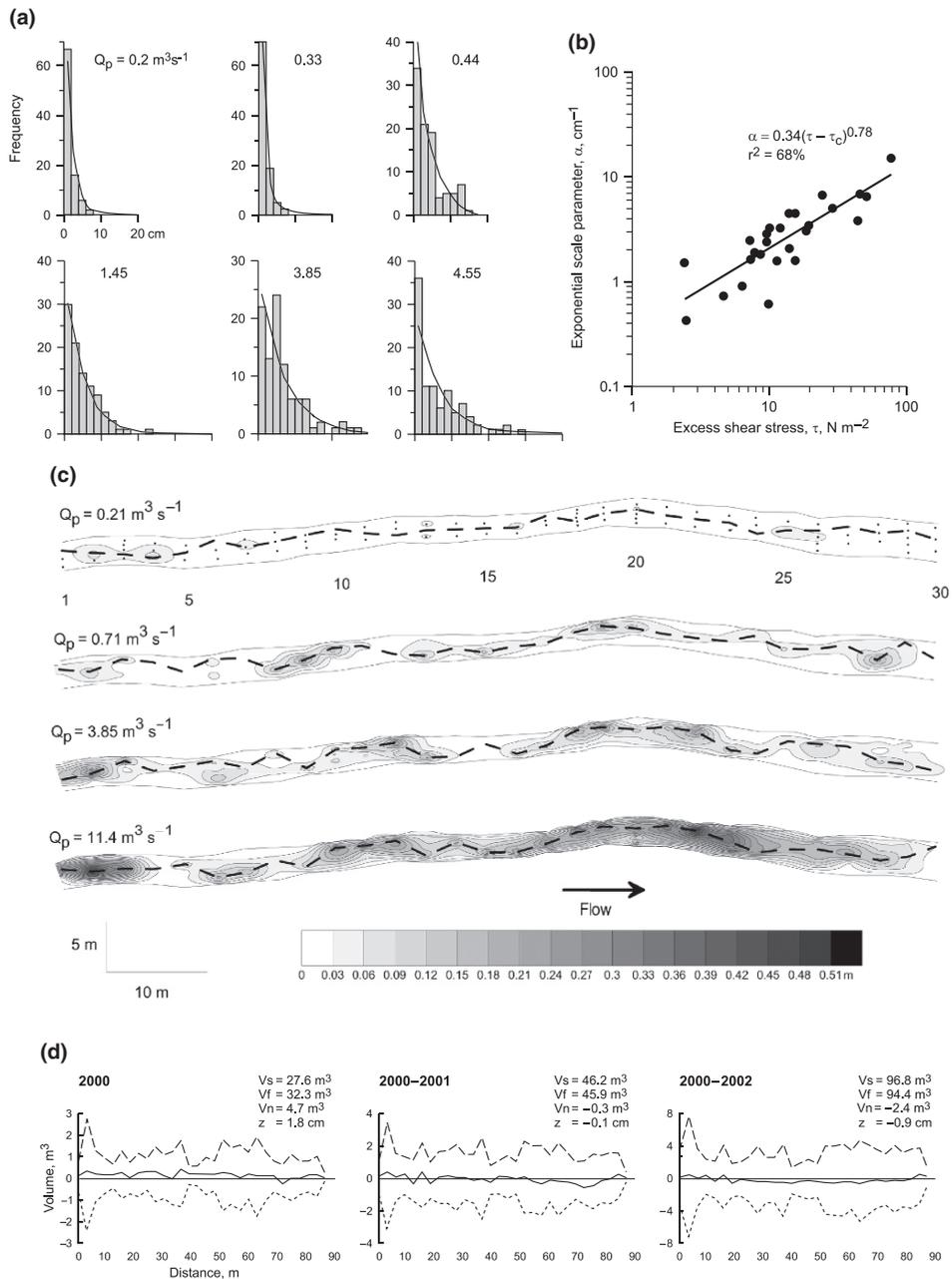


Fig. 12.11 Changes in stream bed elevations during individual events in a low order tributary of Walnut Gulch, Arizona (the main channel of Powell et al. 2005). **(a)** Distributions of scour depths for six events (after Powell et al. 2005). The events are ordered by peak discharge (Q_p ; $\text{m}^3 \text{ s}^{-1}$). Distributions are modelled using the one parameter exponential model. **(b)** Least squares relationship between exponential model parameter and excess shear stress (after Powell et al. 2005). The relationship incorporates data from two additional channels and provides a means to estimate depths of scour in similar streams. **(c)** Spatial

patterns of stream-bed scour and fill for four flow events (after Powell et al. 2006). Cross-section and scour chain locations are shown in the top illustration. The *dashed line* shows the locus of the maximum depth of scour. **(d)** Cumulative patterns of volumetric scour (V_s ; m^3), fill (V_f ; m^3), net change ($V_n = V_s - V_f$; m^3) and average change in stream-bed elevation (z ; m) at the end of three flood seasons (after Powell et al. 2007). Aggradation and degradation fluctuated with no persistent temporal trend so that sediment transfers did not lead to significant and progressive change to the volume of sediment stored within the reach

depths of stream-bed scour provide an attractive and relatively inexpensive means to quantify rates of bed material movement. The method utilises the relation

$$q_{bm} = u_b z_s (1 - p) \rho_s \quad (12.10)$$

where q_{bm} is the mass transport rate of bed material ($\text{kg m}^{-1} \text{s}^{-1}$), u_b is the virtual rate of particle travel (m s^{-1}), z_s and w_s are the active depth and width of the stream bed respectively (m), and p is the porosity of the sediment (Hassan et al. 1992; Haschenburger and Church 1998). Data from the Nahal Yatir in southern Israel (see below) have been used to evaluate the method in a dryland environment. Post-event estimates of bedload yield based on the displacement of gravel-tracers and the depth of scour and fill obtained by scour chains are found to be very similar to that derived from a bedload rating relation derived by direct monitoring (Laronne pers comm.).

Direct Measurement Methods

Direct monitoring during flood events provides numerous opportunities to develop further insights into the dynamics of bedload transport in dryland environments. Of particular significance are the studies undertaken in Nahal Yatir and Nahal Eshtemoa, two neighbouring upland gravel-bed rivers in the Northern Negev Desert, Israel. In these streams, contemporaneous measurements of bedload discharge (q_b ; kg m^{-1}

s^{-1}) and shear stress during flash floods have been obtained using automatic sediment transport monitoring stations comprising a number of Birkbeck-type slot samplers (Reid et al. 1980; Laronne et al. 1992) and stage recorders (Fig. 12.12).

The studies show that the two streams are subject to intense bedload activity (Fig. 12.13a). Maximum recorded channel average transport rates are about $7 \text{ kg m}^{-1} \text{ s}^{-1}$ (Reid et al. 1995, 1998) while rates as high as $12.6 \text{ kg m}^{-1} \text{ s}^{-1}$ are recorded at individual samplers in Nahal Eshtemoa (Powell et al. 1999). The high transport rates reflect the high transport stages ($= \tau/\tau_c$) generated by the flash floods. In Nahal Eshtemoa, for example, all but three of the 19 flow events monitored over a four-year study period generated transport stages of three or more and over 50% generated transport stages greater than five (Powell et al. 2003). As shown in Fig. 12.13a, the relationships between channel-average shear stress and contemporary channel average shear stress for nine events in Nahal Eshtemoa and four events in Nahal Yatir are unusually well defined. Moreover, the predictions of several engineering formulae correspond closely to the observed data suggesting that the measured transport rates approximate the transport capacity of the flow (Reid et al. 1996; Powell et al. 1999). The transport of capacity loads and the simplicity and consistency of the bedload response recorded in these two dryland streams is in marked contrast to that observed in many humid-temperate perennial streams. The differences may be explained by the fact that the beds of Nahal Eshtemoa and Nahal Yatir are not armoured (see

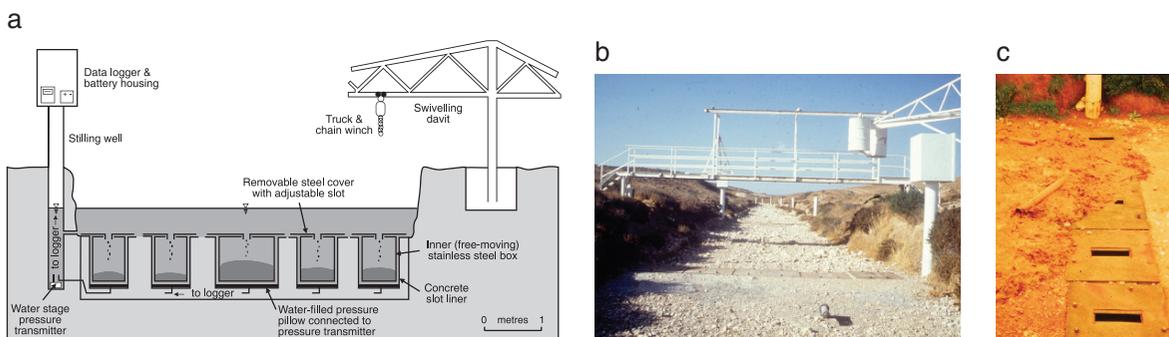


Fig. 12.12 Monitoring bedload transport in Nahal Eshtemoa, Israel. (a) Schematic diagram of the automatic sediment transport monitoring station (after Powell et al. 1999). (b) View upstream through the sediment transport monitoring station (after Powell et al. 2003). Note the five Birkbeck-type bedload samplers in-

stalled across the width of the channel beneath the bridge and the stage recorders extending up the approach reach. (c) Flood bore advancing over the bedload samplers in Nahal Eshtemoa (after Powell et al. 2003)

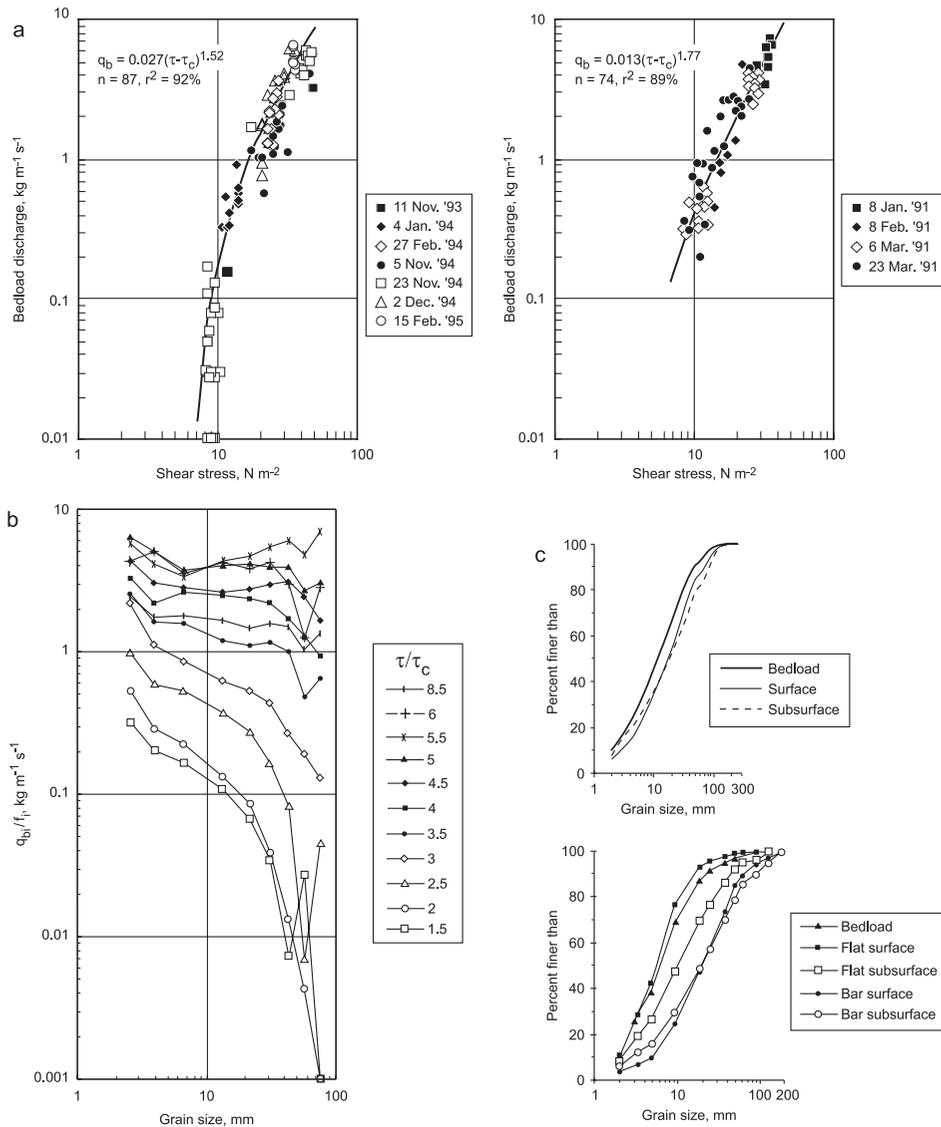


Fig. 12.13 Bedload transport rates and grain size distributions in Nahal Eshtemoa and Nahal Yatir, Israel. **(a)** Channel average bedload transport rates as a function of channel average shear stress in Nahal Eshtemoa (*left*; after Reid et al. 1998) and Nahal Yatir (*right*; after Reid et al. 1995). The curves passing through the data are the ordinary least squares rating relations (the coefficients have been adjusted to eliminate bias due to the log-log transformations). Zero transport rates in Nahal Eshtemoa plotted as $0.01 \text{ kg m}^{-1} \text{ s}^{-1}$. **(b)** Ratio of mean transport rates for the i th size class (q_{bi}) with the frequency of occurrence in the bed (f_i) as a function of the geometric mean grain size (D_{gi}) of each size fraction in Nahal Eshtemoa (after Powell et al. 2001). The near-absence of the coarsest size fractions in the bedload at $\tau/\tau_c < 2$, the over-representation of the finer fractions and the under-representation of coarse fractions at $c. \tau/\tau_c = 3$ and

the equivalence of bedload and bed material grain size distributions at $\tau/\tau_c > 4$ is indicative of partial transport (Wilcock and McARDell 1993), size selective transport (Ashworth et al. 1992) and equal mobility (Parker and Toro-Escobar 2002) respectively. **(c)** Bed material and bedload grain size distributions in Nahal Eshtemoa (*top*; after Powell et al. 2003) and Nahal Yatir (*bottom*; after Reid et al. 1995). The bedload size distribution in Nahal Eshtemoa represents the calibre of the material transported out of the catchment over a four year period estimated using the transport relation of Powell et al. (2001). The bedload size distribution in Nahal Yatir represents the sediment that accumulated in the centre sampler during four events as reported in Laronne et al. (1994). The terms 'bar' and 'flat' refer to contrasting sedimentary units within the reach (see Fig. 12.22a)

below) which reduces well known sedimentological constraints on sediment mobility and availability (Laronne et al. 1994; Reid and Laronne 1995).

It is worth noting that other dryland streams demonstrate more complex bedload responses to changes in flow strength. In Nahal Yael in southern Israel, for example, coarse grained sediment waves were found to migrate through the measuring section every 40–50 min (Fig. 12.14). The origin of the waves is not known, but may be related to catchment and network controls on sediment delivery to the channels. Other workers have highlighted the effect that unsteady flows have on bedload transport rates due to the inability of the bed to adjust as quickly as the flow (e.g. Plate 1994; Lee et al. 2004). The implications of this and related work for sediment transport in flashy dryland streams awaits evaluation.

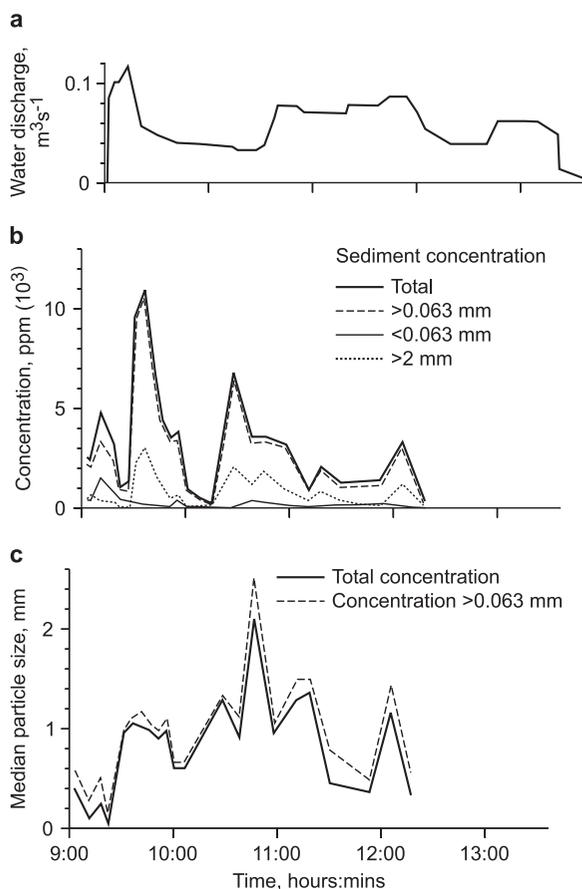


Fig. 12.14 Variation in (a) water discharge, (b) concentration of sediment load and (c) median particle size of sediment load during the event of 20 February 1970 in Nahal Yael, Israel (after Lekach and Schick 1983). Bedload transport occurred as a series of waves that formed independently of the pulses in discharge

In Nahal Eshtemoa, the bedload is fine grained at low flow but coarsens with increasing shear stress, converging with the grain size distribution of the bed at high flows (Fig. 12.13b). The shift in bedload grain size distribution with increasing flow strength accords with the widely held view that transport is partial and size selective at low excess shear stresses but approaches a condition of equal mobility at high levels of excess shear stress (see review by Gomez 1995). Since flow duration increases with decreasing flow magnitude, Wilcock and McArdeell (1997) suggest that partial transport is the dominant transport regime in gravel-bed rivers and results in sediment loads that are considerably finer than the bed material (see also Leopold 1992; Lisle 1995). In Nahal Eshtemoa, partial and size selective transport occurs for 73% of the time the channel is competent to transport bedload. The size distribution of the bedload modelled over a four year period, however, is only slightly finer than that of the bed material (Powell et al. 2001, 2003; Fig. 12.13c). Even though partial and size selective transport conditions dominate and produce bedload size distributions that are finer than the size distribution of the bed material for the majority of the time the stream is geomorphologically active, the rate of transport of the coarser fractions that occurs at high transport stages almost serves to compensate, rendering the size distribution of the annual bedload not that much finer than the bed material. A similar evolution in bedload grain size is observed in Nahal Yatir though the finer bed material ensures that the partial transport domain is largely absent and that the bed is fully mobilised for a greater proportion of time. As a consequence, bed material and bedload grain size distributions also show a close correspondence (Laronne et al. 1994; Fig. 12.13c). A similar dynamic is observed in Goodwin Creek, a seasonal stream in north-central Mississippi (Kuhnle and Willis 1992).

Several authors have questioned whether there are differences in the dynamics of bedload transport between ephemeral and perennial rivers (e.g. Almedeij and Diplas 2003, 2005). Reid et al. (1995) compared bedload transport rates recorded in a number of perennial and ephemeral/seasonal rivers (Fig. 12.15a). They noted that Oak Creek (Oregon, USA; perennial), Turkey Brook (England, UK; perennial) and Nahal Yatir (Israel; ephemeral) define a relatively consistent relation, but that data from East Fork River (Wyoming, USA; perennial), Torlesse Stream (New

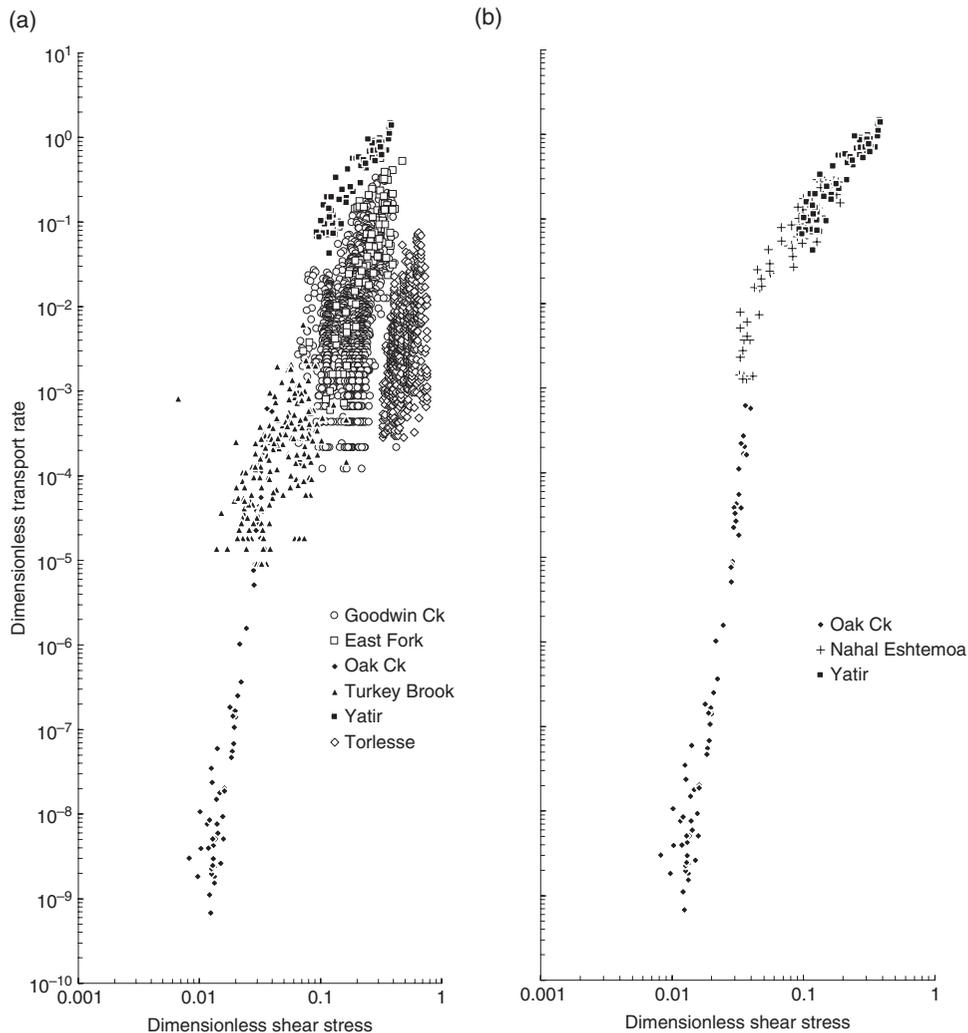


Fig. 12.15 Dimensionless bedload transport rates as a function of dimensionless shear stress in (a) one ephemeral (Yatir), one

seasonal (Goodwin Creek) and four perennial streams (after Reid et al. 1995) and (b) Nahal Yatir, Nahal Eshtemoa and Oak Creek

Zealand; perennial) and Goodwin Creek (Mississippi, USA; seasonal) are shifted to the right, suggesting a different dynamic. However, Goodwin Creek and East Fork River contain significant amounts of sand which can be expected to augment transport rates in a non-linear manner (Wilcock et al. 2001; Wilcock and Crowe 2003). After accounting for the effect of sand on gravel transport rates, Wilcock and Kenworthy (2002) demonstrate that Oak Creek, Goodwin Creek and the East Fork River collapse onto a single curve. The implication of these comparisons is that bedload transport rates measured in perennial and ephemeral rivers fall on different parts of a single continuum that

represents that the bedload-shear stress response of gravel-bed rivers, a conclusion further supported by the fact that the data from Nahal Eshtemoa (Israel; ephemeral) dovetails with the data from Oak Creek, Turkey Brook and Nahal Yatir (Fig. 12.15b). This issue is considered further in the context of stream-bed armours (see below).

Channel Morphology

The morphology of alluvial channels develops through spatially and temporally variable patterns of erosion,

transport and deposition. Much of the research on alluvial channel forms has been conducted in humid-temperate rivers and is based on the identification and analysis of equilibrium channel forms. Four aspects of channel morphology are usually considered: (i) the shape and size of the channel cross-section; (ii) the configuration of the channel bed; (iii) the river longitudinal profile and slope and (iv) the channel pattern. This conceptual framework is adopted here though it is recognised that our ability to make rational generalisations about dryland river forms is hampered by that fact that many dryland rivers fail to exhibit equilibrium behaviour.

Channel Equilibrium and Formative Events

Equilibrium concepts are relevant to medium timescales over which, it is reasoned, rivers develop a relatively stable and characteristic morphology that allows them to transmit the imposed water and sediment discharges (Mackin 1948; Leopold and Bull 1979). Explanations for the form of channels in equilibrium are usually sought in terms of a single 'dominant' or 'formative' discharge, a statistically- or morphologically-based construct that replaces the frequency distribution of flows. Wolman and Miller (1960) defined the dominant discharge as the flow which cumulatively transports the most sediment. They argued that the geomorphological effectiveness of a particular discharge magnitude is the product of

the sediment transported by an event of that magnitude and its frequency of occurrence. Using a sediment transport law and flood frequency distribution parameterised for humid-temperate conditions, they demonstrated that the most effective flood is defined by an event of moderate magnitude and frequency (Fig. 12.16a). Other workers have defined dominant discharge in terms of the flow that determines particular channel parameters such as the cross-sectional capacity of the flow (Wolman and Leopold 1957) or the wavelength of meander bends (Ackers and Charlton 1970).

The extensive debate that surrounds the concepts of dominant discharge and equilibrium adjustment is beyond the scope of this review (see Phillips (1992) and other papers from the 23rd Binghampton Symposium; Thorn and Welford 1994; Bracken and Wainwright 2006). It is worth noting, however, that the explanatory power of the two concepts in dryland environments is often questioned. The hydrological regime of many dryland rivers generates large differences between high and low flows and pronounced spatial and temporal discontinuities in process operation which makes the definition of formative discharges and the recognition of equilibrium forms difficult (Thornes 1980; Schick et al. 1987; Bourne and Pickup 1999; Hooke and Mant 2000; Coppus and Imeson 2002). Moreover, many dryland rivers appear not to exhibit equilibrium behaviour. As explained below, this contrast between dryland and humid-temperate river behaviour is due to fundamental differences in magnitude-frequency relationships (Baker 1977) and relaxation times (Wolman and Gerson 1978; Brunsten and Thornes 1979).

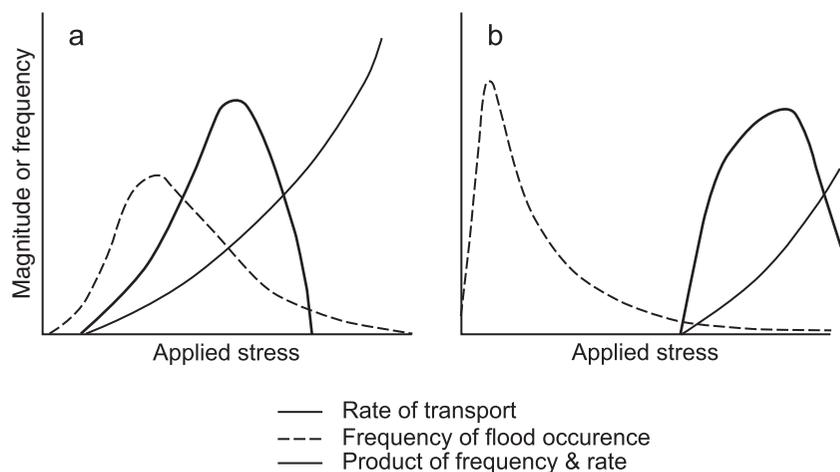


Fig. 12.16 (a) Generalised magnitude-frequency relationships of Wolman and Miller (1960). (b) Modifications to the magnitude-frequency relationships of Wolman and Miller (1960). After Baker (1977)

Baker (1977) re-evaluated Wolman and Miller's (1960; Fig. 12.16a) magnitude-frequency analysis in a dryland context by arguing that in dryland environments, the mode of the flood frequency distribution shifts left (because the ratio of high to small flow events is larger) and that the sediment transport law shifts to the right (because the sediments are generally coarser and have higher entrainment thresholds). The resultant increase in the magnitude and decrease in the frequency of the flow that transports the most sediment implies that the rare catastrophic event is more important in shaping dryland streams (Fig. 12.16b). Moreover, Wolman and Gerson (1978) recognised that the geomorphological effectiveness of high magnitude flows is further enhanced in dryland rivers by the limited occurrence of high frequency/low magnitude flows and the absence of sediment-trapping vegetation that facilitates channel recovery in humid-temperate environments (Fig. 12.17a-c). Although active channels may show some short-term adjustment to the prevailing hydrological regime, these equilibrium channels are often superimposed on a palimpsest disequilibrium morphology produced by more infrequent, higher magnitude events (Rhoads 1990) while in hyper-arid environments, channel recovery may be virtually non-existent such that successively larger floods leave permanent imprints on the landscape (Fig. 12.17d).

These climatically controlled contrasts in river behaviour can be characterised by the transient form ratio (TF) defined as the ratio of the mean relaxation time to the mean recurrence interval of significant channel disturbing events (Brunsden and Thornes 1979). Fluvial systems for which $TF < 1$ can develop equilibrium channel forms because they adjust to new conditions or recover from flood-induced channel change before the next major disturbance occurs (Fig. 12.17a,b). Many dryland rivers, however, may be characterised by $TF > 1$ with the result that channel forms display either dis-equilibrium or non-equilibrium behaviour (Stevens et al. 1975; Rhoads 1990; Bourne and Pickup; 1999; Beyer 2006). As defined by Renwick, (1992), the former represents instances when the development of an equilibrium state is precluded by long relaxation times (Fig. 12.17c) while the latter occurs when systems display no net tendency toward an equilibrium state (Fig. 12.17d).

Tooth and Nanson (2000a) argue that non- or dis-equilibrium channel behaviour is not a character-

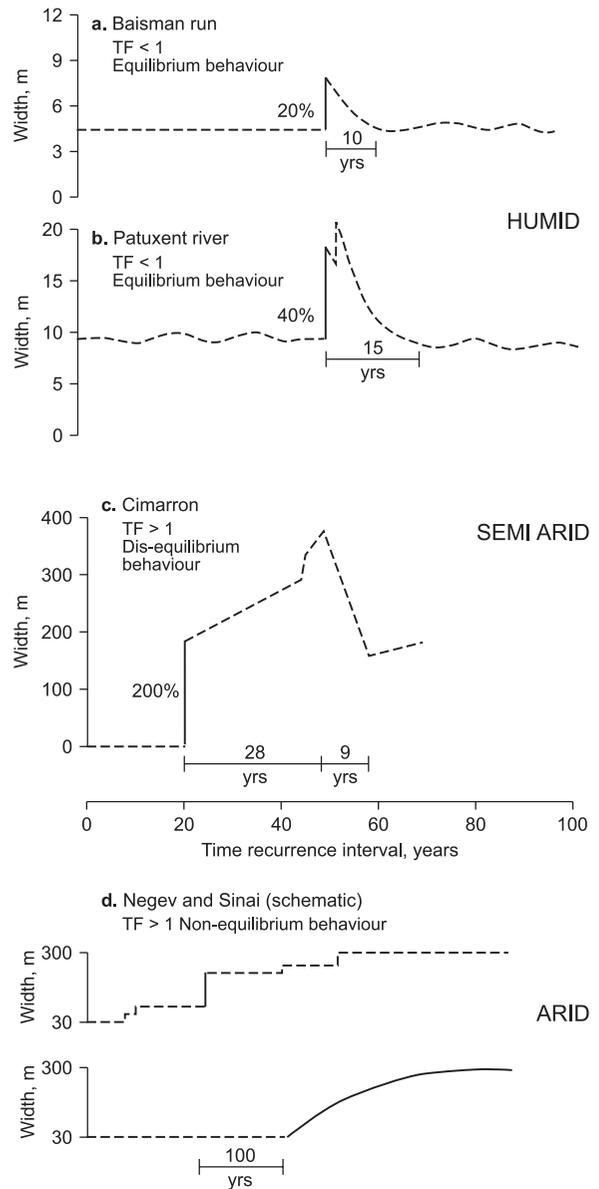


Fig. 12.17 Temporal changes in channel width in rivers from different hydro-climatic regimes showing relaxation (recovery) times following storm disturbances (after Wolman and Gerson 1978). Channel recovery following major flood disturbances occurs in response to sedimentation during frequent low-magnitude restorative flows and occurs most rapidly and effectively in humid-temperate environments. TF is the transient form ratio of Brunsden and Thornes (1979). Equilibrium terminology after Renwick (1992)

istic of all dryland rivers. They suggest that that the potential for dryland rivers to develop an equilibrium channel form is a function of local hydrological, geomorphological and sedimentological conditions.

High energy environments with low erosion thresholds (steep, low-order rivers subject to short-lived high magnitude flash floods carrying large amounts of coarse bedload) favour the development of non- (dis-) equilibrium channels whilst lower energy environments with higher erosion thresholds (medium – large low gradient rivers with resistant, confining banks subject to long duration floods) favour the development of equilibrium forms. The latter conditions typify the medium-large sized catchments of the Northern Plains and Channel Country of central Australia where channels meet several criteria said to characterise equilibrium conditions (stability despite the occurrence of large floods, sediment transport continuity, strong correlations between channel form and process; an adjustment to maximise sediment transport efficiency).

Cross-Sectional Form

The dominant control on the cross-sectional dimensions of a river is discharge. This is perhaps best illustrated by Ferguson's (1986) observation that channel width and depth increase systematically with increasing bankfull discharge as it varies over nine orders of magnitude from small laboratory channels to the world's largest rivers. Empirical geomorphological investigations of the relationships between channel geometry and stream discharge have traditionally followed the downstream hydraulic geometry approach of Leopold and Maddock, (1953) in which downstream changes in width (w ; m), depth (y ; m) and velocity (u ; $m\ s^{-1}$) are expressed as power functions of an assumed dominant discharge (a discharge at a specified frequency of occurrence (Q_x ; $m^3\ s^{-1}$):

$$w = aQ_x^b \quad (12.11)$$

$$y = cQ_x^f \quad (12.12)$$

$$u = kQ_x^m \quad (12.13)$$

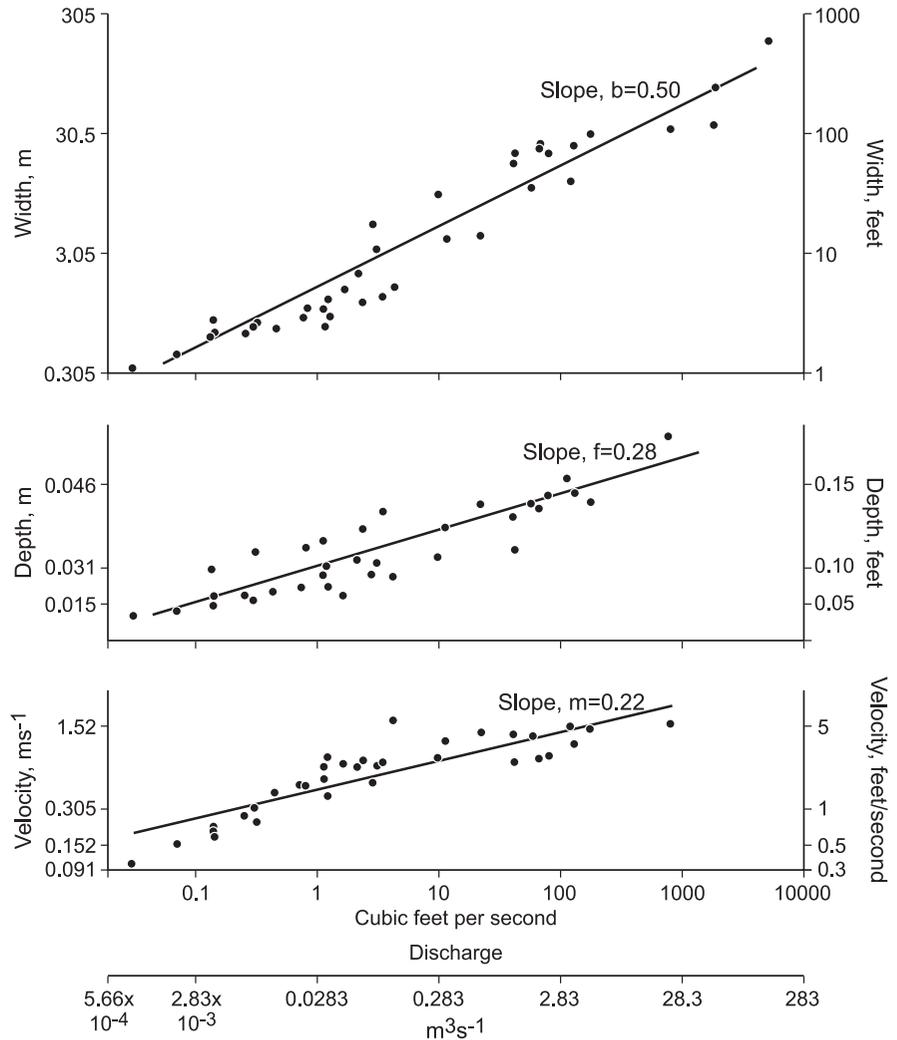
The exponents $b = 0.5$, $f = 0.4$ and $m = 0.1$ defined for streams in the American Midwest using the mean annual flood are often used to characterise the downstream adjustment of humid temperate perennial streams to increasing discharge. The exponent set indicates that width increases faster than depth (gener-

ating downstream changes in channel shape as indexed by the width:depth ratio) and that velocity increases downstream (contradicting traditional Davisian assumptions). Although comparative data from dryland environments are sparse, compilations of hydraulic geometry exponents suggest some regional variation according to climatic regime (Park 1977; ASCE 1982). A study of the downstream adjustment of ephemeral channels in New Mexico, USA shows that although the increase in width is about the same as that observed in humid-temperate perennial rivers, the increase in velocity is more rapid and the increase in depth is less rapid (Leopold and Miller 1956; Fig. 12.18). The different response of the ephemeral channels is attributed to a downstream increase in suspended sediment concentrations that decreases turbulence and bed erosion.

The erodibility of channel banks exerts important secondary controls on cross-sectional adjustment. Since dryland weathering processes do not produce significant amounts of cohesive silts and clays, bank materials often lack the strength to resist processes of bank erosion. As a result, channels tend to respond to floods by widening, rather than deepening, their cross-section. Schumm (1960), for example, showed that width:depth ratios are negatively correlated with the silt-clay content of perimeter sediments (an index of bank shear strength and erodibility; Fig. 12.19).

Merritt and Wohl (2003) examined the downstream adjustment of Yuma Wash in SW Arizona to an event with a discharge estimated at c. 20% of the maximum probable flood. They found that increases in width were substantial ($b = 0.78$) whereas the increases in depth and velocity were modest ($f = 0.15$ and $m = 0.14$). They attributed the rapid increase in channel width to the low cohesion of the bank material which comprised less than 3% silt and clay. The behaviour of Yuma wash contrasts with channels in the northern Negev Desert which are able to maintain relatively deep and narrow cross-sections despite the high transport stages generated by flash floods (Laronne and Reid 1993; Reid et al. 1998; Fig. 12.12b). The absence of significant bank retreat in these channels has been attributed to the cohesive properties of the loess-rich soils (Powell et al. 2003). In non-cohesive sediments, flood-induced increases of channel width can be dramatic. Schumm and Lichty (1963) describe how a major flood in 1914

Fig. 12.18 Downstream hydraulic geometry relations of ephemeral channels in New Mexico (after Leopold and Miller 1956). Data were collected at discharges of unknown frequency



in SW Kansas began a process of channel widening that increased the average width of the Cimarron River from 15 m in 1874 to 366 m by 1939. Similar transformations of channel width have been observed in other sand-bed rivers of the SW USA including Plum Creek, a tributary of the South Platte River, Colorado (Osterkamp and Costa 1987), the Santa Cruz River in Arizona (Parker 1993) and the Gila River in SE Arizona (Burkham 1972; Huckleberry 1994; Hooke 1996; Fig. 12.20).

Bank stability is also controlled by vegetation. Although vegetation is generally sparse in semi-arid environments, many dryland rivers support dense stands of riparian vegetation which influence bank stability through mechanical and hydrological effects (Simon and Collinson 2002; Simon et al. 2004). The latter are

likely to be particularly important in semi-arid environments where banks are typically unsaturated and susceptible to changes in soil moisture levels (Katra et al. 2007). Stabilising effects include root-binding of sediment which increases the tensile strength and elasticity of soils and helps to distribute shear stresses rather like the bars in reinforced concrete or the fibres in a carbon fibre material (Tal et al. 2003) and enhanced canopy interception and evapotranspiration which results in better drained bank materials with reduced bulk weight and lower positive porewater pressures. Riparian vegetation also increases flow resistance, thereby decreasing flow velocities and the shear stress available for erosion (Thornes 1990; Wilson et al. 2005). Destabilising effects of vegetation include bank loading by the weight of trees and higher near-surface moisture

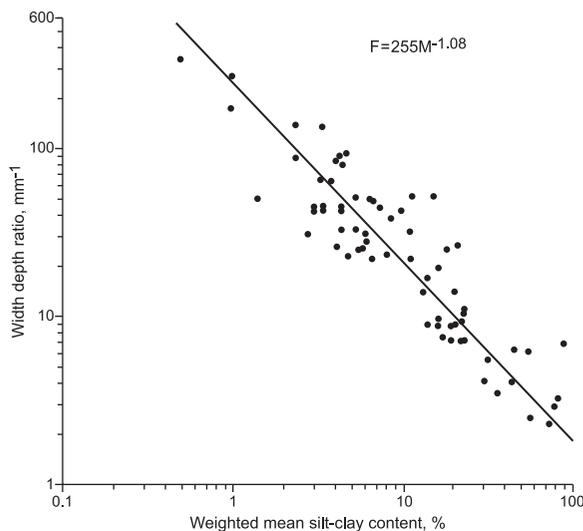


Fig. 12.19 Relation between width:depth ratio (F) and weighted mean percent silt-clay content in the channel boundary (M ; after Schumm 1960)

contents after rainfall due to increased soil infiltration capacities. Vegetation plays an important role in narrowing channels after they have been widening by major flood events. The effects of vegetation on channel recovery are discussed below in the context of channel planform adjustment.

In a spatial context, Wolman and Gerson (1978) demonstrate that the rate of change of channel width with increasing drainage is more rapid in dryland rivers than it is in humid-temperate rivers, at least in catchments up to about 100 km^2 (Fig. 12.21a). Interestingly, dryland channels draining larger catchments maintain near constant widths. This has been attributed to various factors including an imposition of an upper limit on stream discharge caused by the limited areal extent of storm events (Sharon 1972, 1981; Renard et al. 1993; Goodrich et al. 1995) and/or transmission losses (see Chapter 11) and in hyper-arid environments, the lack of channel recovery between events (Fig. 12.17d). Where transmission losses exceed tributary inflows, the resultant downstream decrease in discharge downstream can lead to concomitant reductions in channel width and depth (e.g. Dunkerley 1992; Fig. 12.21b) leading to the termination of channelised flow and bedload transport in broad low gradient surfaces known as floodouts (Tooth 1999; Fig. 12.21c). It is not known whether the hydraulic geometry exponents that model the downstream

increase in channel dimensions under conditions of increasing discharge (Fig. 12.18) also describes the downstream decrease in channel dimensions observed under conditions of decreasing discharge.

The complexity of channel width adjustment in large dryland rivers is demonstrated by Tooth (2000b) who documented changes in channel character along the length of the Sandover, Bunday (Sandover-Bunday) and Woodforde Rivers in central Australia (Fig. 12.20c). In all three rivers, distinct form-process associations define four contrasting fluvial environments: confined upland, piedmont, lowland zones and unconfined floodout zones. Channel widths tend to increase throughout the upland and piedmont zones where integrating channel networks cause discharge to increase downstream. In the lowland zone, transmission losses exceed tributary recharge and widths and depths decrease downstream until the flows and sediments dissipate in the floodout zone. Although the rivers exhibit variable patterns of downstream channel change and several unusual channel characteristics (e.g. anabranching and aggrading floodout zones), Tooth (2000b) concludes from a qualitative review of channel pattern parameters that many aspects of channel form (including channel width) are 'strongly correlated to and sensitively adjusted to tributary inputs of water and sediment' (Tooth 2000b p. 200).

Where systematic relationships between cross-section channel geometry and discharge exist, they suggest a functional adjustment of channel form to the imposed discharge, the nature of which should be amenable to rational explanation using hydraulic and sediment transport principles. The development of deterministic solutions for the geometry of river cross-sections is hampered by the fact that the degrees of freedom for alluvial channel adjustment exceed the number of available equations. The traditional approach is to assume that width, depth, velocity and either slope or sediment load adjust to the other of these two variables plus discharge and grain size (Ferguson 1986) such that a solution is provided by solving the flow continuity relationship, a flow resistance law, a sediment transport equation and assuming either (i) a threshold channel, (ii) maximum efficiency criterion in conjunction with a bank stability criterion or (iii) by fitting an empirical relation to one variable (see Ferguson 1986 for a review of approaches). The relative merits of the different approaches are subject to some debate in part, because they all fail to account

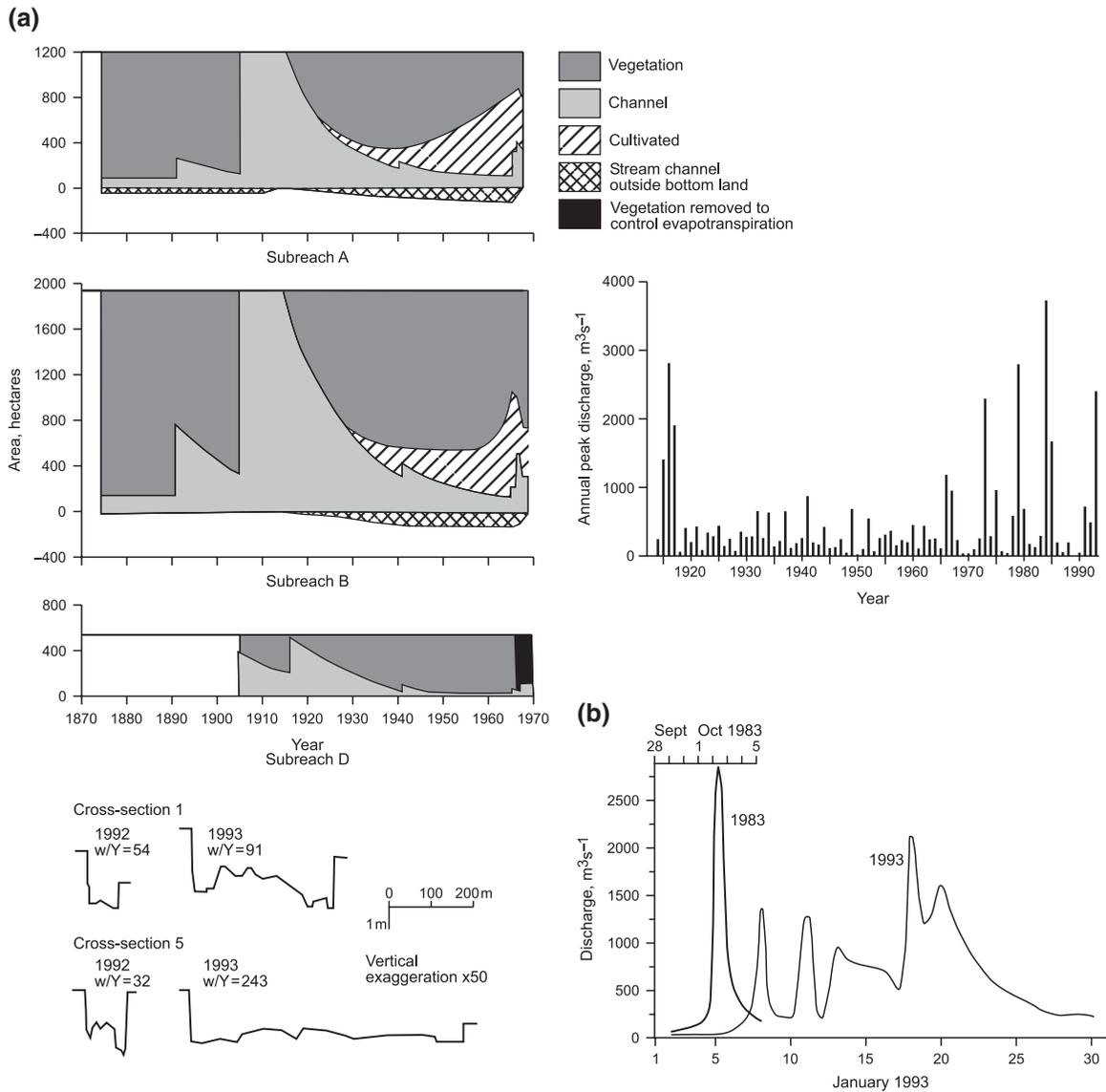


Fig. 12.20 Discharge records and changes in channel width in the Gila River, Arizona, USA (a) 1875–1968 near Safford (after Burkhams 1972) and (b) 1993–1993 near Florence (after

Huckleberry 1994). Note that the channel did not widen appreciably during the 1983 flood

for a variety of real-world complications (Eaton and Millar 2004; Millar 2005). However, they all have the same qualitative outcome in which the steady-state morphology is associated with a characteristic value of dimensionless shear stress (Ferguson 1986). The explanatory power of the approach has yet to be tested in a dryland river showing regularity in cross-section adjustment.

There remains considerable uncertainty as to how channels adjust their cross-sections. Much is known

about the geotechnical and hydraulic forces that control bank stability and retreat, and attempts have been made to couple models of specific bank erosion processes (fluvial entrainment and mass wasting of bank materials and the downstream transport of failed bank materials) to predict cross-section adjustment in alluvial channels (e.g. Simon et al. 2000). Further work, however, is needed better to understand how hydraulic and gravitational processes interact to control rates of bank retreat and channel widening and to incorporate

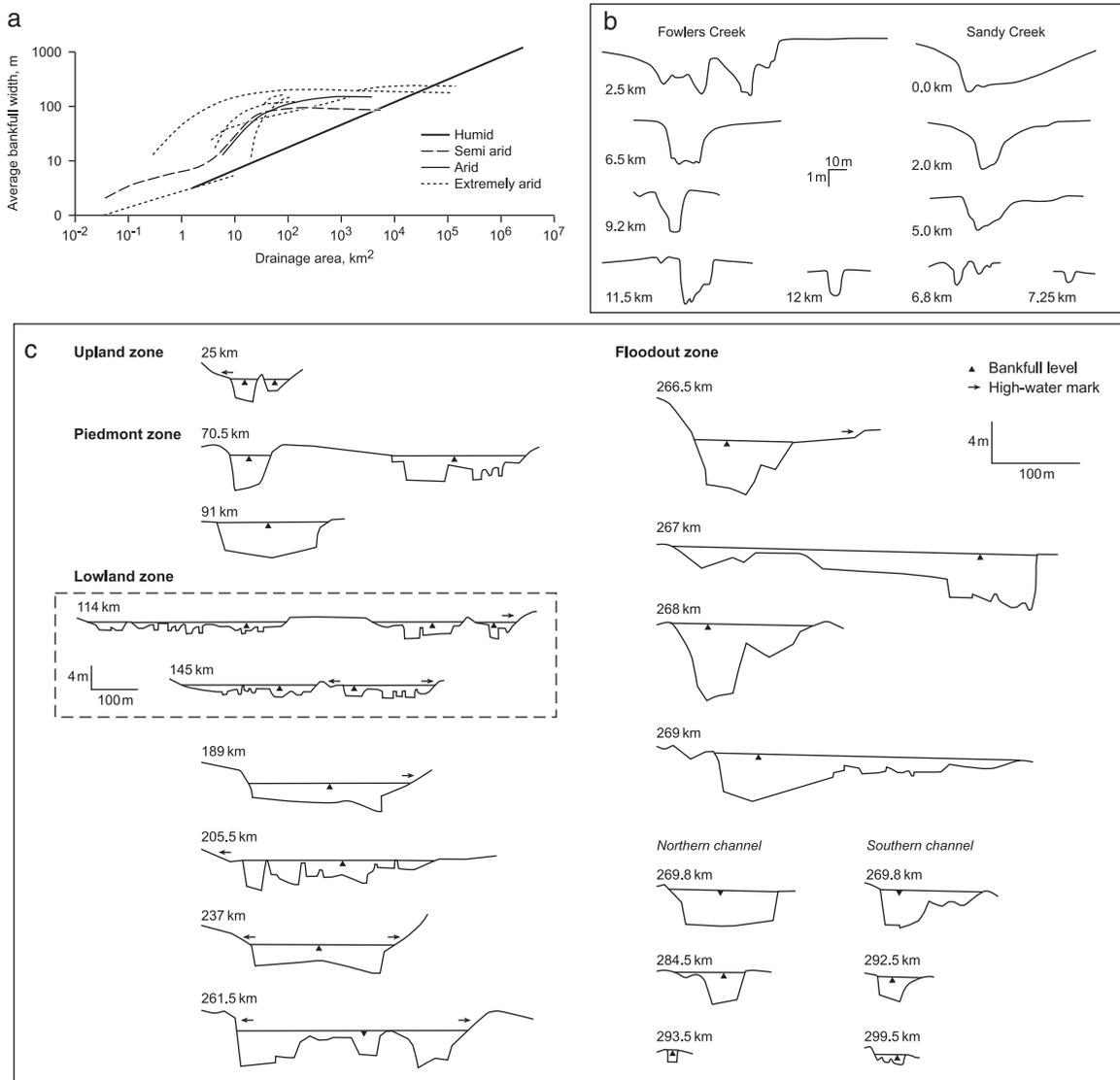


Fig. 12.21 (a) Changes in channel bankfull width with increasing drainage area in different climatic settings (after Wolman and Gerson 1978). (b) Downstream changes in channel width in Fowler and Sandy Creeks, western NSW, Australia (after

Dunkerley 1992). (c) Downstream changes in channel width in the Bunday (Sandover-Bunday) River in the northern plains of central Australia (after Tooth 2000b)

this understanding into existing models of flow, sediment transport and morphological change (e.g. Darby et al. 2002).

Bed Configuration and Texture

Bed configuration and texture represent two of the most adjustable components of channel form with

potential for regulating the short-term and mutual adjustment of water flow, sediment supply and grain size at a range of spatial and temporal scales. The complex relationships between flow, sediment transport and bedform geometry that facilitate this adjustment are beyond the scope of this review. Comment is therefore restricted to some important aspects of the bed morphology and sedimentology of dryland streams.

Bed Configuration

In humid-temperate environments, single-thread channels with coarse heterogeneous sediments on low-moderate slopes ($< 2\%$) often develop an alternating pattern of coarse-grained topographic highs (riffles) and finer-grained topographic lows (pools) with a wavelength of about 5–7 channel widths (Keller and Melhorn 1978). On steeper slopes, the bedform evolves to a step-pool sequence with a wavelength of about 2 channel widths (Chin 2002). Riffles and steps are significant sources of flow resistance that concentrate energy losses at particular locations along the course of a river (Church and Jones 1982; Abrahams et al. 1995; Chin and Phillips 2007). Since the development of riffles-pools and steps-pools reflects a significant aspect of channel adjustment, they are widely regarded as equilibrium channel forms.

The undulating topography of the pool-riffle sequence is conspicuously absent from many single-thread dryland rivers though their sediments still appear to be distributed in patterns associated with a typical sequence. Reid et al. (1995) for example, describe how the bed material of Nahal Yatir, a gravel-bed river in the Northern Negev Desert, is characterised by an alternating pattern of comparatively coarse ‘bars’ ($D_{50} = 20\text{ mm}$) and longer, planar, finer ‘flats’ ($D_{50} = 6\text{ mm}$; Fig. 12.22a). These bedforms have little or no topographic expression and their positions are stable over time, despite the

passage of competent floods. The neighbouring Nahal Hebron has a comparable sedimentology comprising a ‘barely discernible alternation of gravel bars and granular-sandy pools’ (Hassan 1993, p. 109). Similar patterns of sediment sorting appear in mixtures of sand and gravel. Local concentrations of gravel on otherwise planar, sandy beds have, for example, been described in the arroyos of northern New Mexico (Leopold et al. 1966, Fig. 151) and in the channels of the East Rudolf sedimentary basin in Northern Kenya (Frostick and Reid 1977, p. 2). Comparable alternating sequences of coarser and finer sediments have been identified in steeper channels that might have been expected to form steps and pools (Bowman 1977, Fig. 12.22b).

Intriguingly, the gravel accumulations of Leopold et al. (1966) have a spacing of five-seven times the channel width and are likened to riffles that formed as a kinematic wave. Little was known about steps and pools at the time of Bowman’s (1977) work, but his descriptions indicate that although the channels lack the stair-case morphology of a conventional step-pool system, the coarser segments share many other characteristics including a wavelength of about twice the channel width, steep gradients, high roughness, gravelly-bouldery composition and an association with infrequent discharges and near- or super-critical flow (Montgomery and Buffington 1997). These observations suggest that single-thread ephemeral stream channels develop distinct patterns of longitudinal sediment sorting that are analogous to the pool-riffle

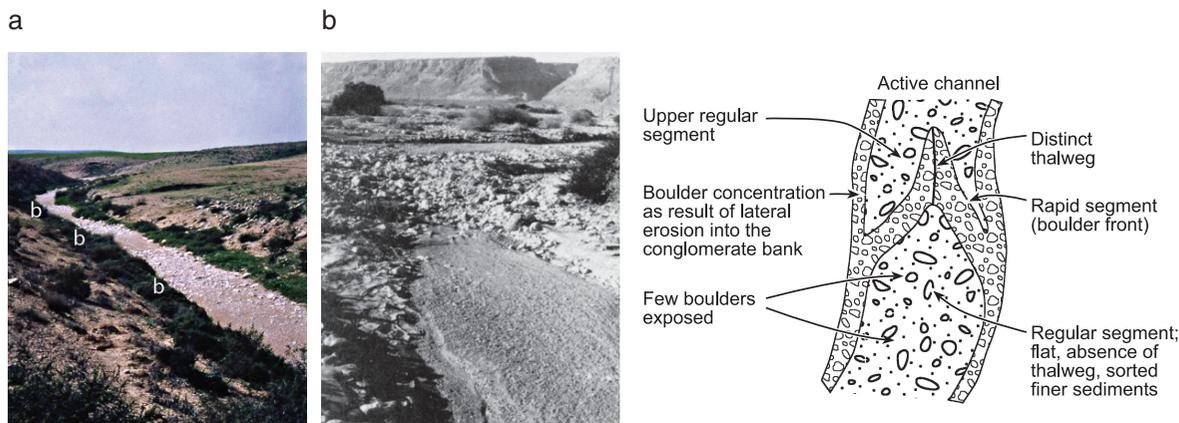


Fig. 12.22 Sedimentary units in dryland streams in southern Israel. (a) Longitudinal alternation of coarse channel bars (‘b’ placed on the adjacent channel bank) and finer ‘flats’ in Nahal

Yatir (after Reid et al. 1995). (b) Stepped bed morphology in Nahal Zeelim (after Bowman 1977)

and step-pool sequences of their perennial counterparts. However, the sedimentary characteristics and dynamics of semi-arid fluvial systems have not been widely studied and the extent to which dryland rivers develop symmetrically repeating bed configurations, the origin, form and function of which can be likened to, or distinguished from, the continuum of channel morphologies identified in humid-temperate streams (e.g. Montgomery and Buffington 1997) is not known.

Small-Scale Bedforms in Sand- and Gravel-Bed Channels

Alluvial rivers develop a wide range of smaller scale bedforms that provide additional sources of flow resistance and also reduce particle mobility. They include ripples, dunes, and antidunes in sand-sized sediments and particle clusters, transverse ribs and bedload sheets in gravels and in mixtures of sands and gravels (Allen 1982; Best 1996). The occurrence of different bedforms is usually determined using bedform phase diagrams. These are largely based on laboratory experiments, heavily biased towards sand-sized sediments and define equilibrium bedform regimes in terms of sediment mobility (grain size or fall velocity) and flow intensity (velocity, shear stress, stream power). Such diagrams need to be applied with caution in dryland streams for two reasons. First, research has shown that bedforms in sand-bed rivers have a minimum relaxation time in which they are able to respond to changes in flow conditions (Simons and Richardson 1963; Allen 1973). The implication of this work is that equilibrium bedforms are unlikely to develop in dryland rivers with flashy discharge regimes (Jones 1977). Second, most research conducted on gravel bedforms is either conducted under, or guided by, field conditions observed in humid-temperate rivers in which the majority of sediment transport occurs under the regime of partial transport during perennial flows (Hassan and Reid 1990; Hassan and Church 2000; Wittenberg and Newson 2005; Oldmeadow and Church 2006). Compared to our understanding of sediment sorting and the autogenetic modification of bed surface grain size (see below) very little is known about the structural sedimentology of gravelly beds, and the extent to which the ephemeral discharge regime and the high rates of sediment

supply and transport restricts the development of bed structures in dryland streams remains to be assessed. In this context, it is interesting to note that Hassan (2005) found little evidence of imbrication or other surface structures such as stone cells or particle clusters on the surfaces of channel bars in Nahal Zin in the Negev Desert, Israel. Marked contrasts in the structural sedimentology of dryland ephemeral and humid-temperate perennial rivers were also recorded by Wittenberg (2002).

Bed Texture Adjustment in Gravel-Bed Rivers

Most gravel-bed rivers develop a coarse surface armour layer that overlies finer subsurface material (Fig. 12.23a). Several workers, however, have noted that the surface and subsurface sediments of many dryland, gravel-bed rivers are not markedly different (Schick et al. 1987; Laronne et al. 1994; Hassan 2005; Laronne and Shlomi 2007; Hassan et al. 2006, Fig. 12.23b). The weakly- or un-armoured nature of alluvial gravels in dryland environments has been attributed to high rates of sediment supply and bedload discharge, the limited duration of flash flood recession limbs and, of course, the absence of baseflow (Laronne et al. 1994). Some of these issues are explored below.

Since the surface of a gravel-bed stream becomes finer with increasing transport stage, eventually approaching the grain size of the substrate in the limit of large τ/τ_c (Andrews and Parker 1987), the weak armouring of many dryland channels is consistent with the high transport stages they sustain. Parker (2008) illustrates the dynamic with reference to the unarmoured Nahal Yatir and Sagehen Creek, a perennial stream in the Sierra Nevada of California with a well armoured bed. In the following discussion, f_{ui} , D_{ug} , D_{b50} , D_{bg} , τ_{50}^* and τ_{c50}^* denote the fraction of subsurface material in the i th grain size range, the subsurface mean grain size (m), the median bedload grain size (m), the mean bedload grain size (m), the dimensionless shear stress for D_{50} and the critical dimensionless shear stress for D_{50} respectively. The analysis uses ACRONYM2 of Parker (1990a,b) to predict the surface size distribution and shear velocity required to transport a given bedload size distribution and transport rate. The input bedload size distribution for both streams was ap-



Fig. 12.23 Contrasts in surface armouring between (a) River Wharfe (humid-temperate perennial river, England, UK) and (b) Nahal Yatir (dryland ephemeral river, Israel)

proximated by the subsurface size distribution of Sagehen Creek since in the normalised form of f_{ui} versus D_{gi}/D_{ug} , the grain size distribution also approximates that of Nahal Yatir. The simulation was conducted using a range of transport rates. The predicted values of the ratio D_{50}/D_{b50} and the predicted grain size distributions of the static and mobile armours at different values of τ_{50}^* are shown in Fig. 12.24, together with estimates of τ^* at bankfull flows in Sagehan Creek and at low and high flows in Nahal Yatir. D_{50}/D_{b50} ratios in Sagehan Creek show little change with increasing dimensionless shear stress and the mobile armour is considerably coarser than the bedload at bankfull dimensionless shear stresses ($\tau^* = 0.059$). In contrast, the mobile armour in the Yatir has become much closer to the bedload than the static armor at $\tau_{50}^* = 0.1$ and the armouring has vanished relative to the bedload at $\tau_{50}^* = 0.3$. The evolution and convergence of the surface grain size distribution from a static armour at low

transport rates (when $\tau_{50}^* \approx \tau_{c50}^*$) to that of the bedload at very high transport rates (when $\tau_{50}^* \gg \tau_{c50}^*$) is clearly shown in Fig. 12.24b. It is worth highlighting that the model assumes that the bedload size distribution is constant and approximates the subsurface size distribution. Parker (2008) notes that this is generally not the case (e.g. Fig. 12.13b) and that the bedload grain size dependence on shear stress will reduce the convergence of bedload and bed surface size distributions. Nevertheless, the clear implication is that the lack of armour in dryland streams represents a dynamic sedimentological response to high transport rates generated by high dimensionless shear stresses.

This explanation for the lack of a coarse surface layer dryland river gravels is based on a view of sediment transport dynamics that regards armouring to be a natural consequence of the transport of sediment mixtures at values of shear stresses that prevail in most gravel-bed rivers. Dietrich et al. (1989), however, view

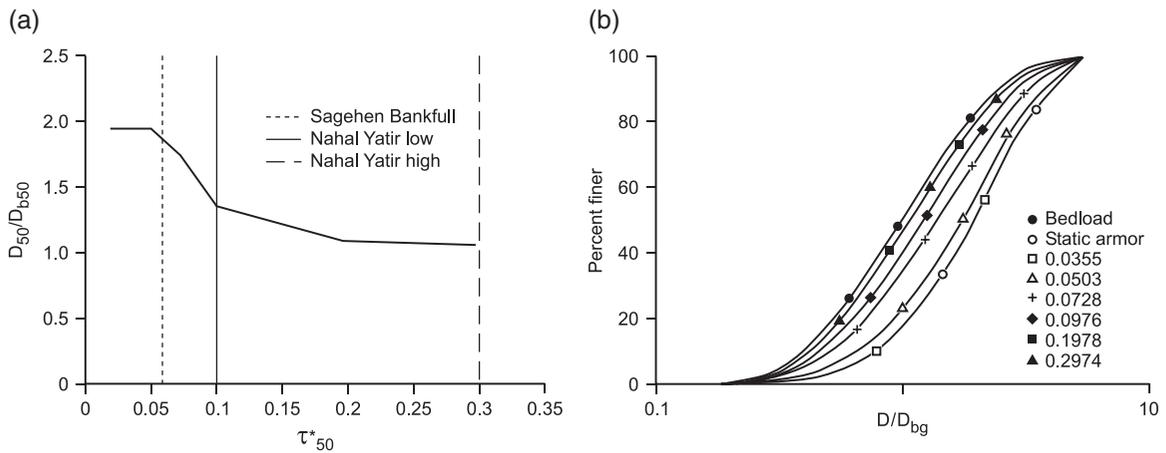


Fig. 12.24 (a) Modelled variation in ratio of median surface grain size (D_{50}) to median bedload grain size (D_{b50}) with τ_{50}^* . (b) Normalised bedload grain size distribution (assumed) and pre-

dicted size distributions for static and mobile armours at values of τ_{50}^* shown in the legend. After Parker (2008)

surface coarsening to be related to the balance between rates of sediment supply and transport. They argue that transport rates in excess of sediment supply induce selective erosion and surface coarsening (e.g. as seen downstream of dams in regulated rivers) while a surplus of sediment forces the deposition of the finer fractions and a consequent reduction in surface grain size (Lisle and Hilton 1992). As a result, surface grain size can be used as an indicator of sediment supply (Buffington and Montgomery 1999). From this perspective, the near equivalence of surface, subsurface and bedload size distributions in dryland channels reflects the high rates of sediment supply from the sparsely vegetated hillslopes. Data collected by Hassan et al. (2006) from arid and humid-temperate streams with a range of sediment supply regimes confirms that many dryland streams are weakly armoured and that sediment supply is a first order control on surface texture and the development of a coarse surface layer. They also suggest that hydrograph shape plays a secondary role. Flume experiments conducted to investigate the influence of hydrograph characteristics on the development of channel armours demonstrate the importance of flow duration and hydrograph symmetry for the development of armoured surfaces.

Channel Pattern

Although rivers exhibit spatial and temporal transitions in channel pattern in response to variations in the

magnitude, frequency and sequencing of flood events, they are often classified by their planform geometry into single-thread (straight and meandering) or multi-thread (braided and anabranching) forms. Straight and anabranching channels are relatively uncommon which suggests that they develop under a restricted set of environmental conditions. The former, for example, tend only to occur in locations where channel alignment is forced by geological controls as illustrated by the drainage of Walnut Gulch in SE Arizona (Murphy et al. 1972). Walnut Gulch and its tributaries drain alluvial fills of Tertiary and Quaternary age and are characterised by wide, shallow, sinuous single- and multi-thread courses. However, the channels are essentially straight where they are bounded by outcrops of caliche and traverse resistant conglomerates with marked and abrupt changes of direction signalling entrenched, fault-controlled drainage. Meandering is the most frequently occurring channel planform at the global scale (Knighton 1998, p. 231). Within dryland environments, however, braided channels are more common than meandering channels (Graf 1988, p. 201). Braided rivers are characterised by frequent shifts in channel position and those in drylands are no exception (Graf 1981, 1983). As a result, the braided channel form has often been regarded as disequilibrium aggradational response to high sediment loads. However, even though individual channels may be transient, the fact that braiding appears to be favoured by particular environmental conditions (high and variable discharges, steep slopes, dominant

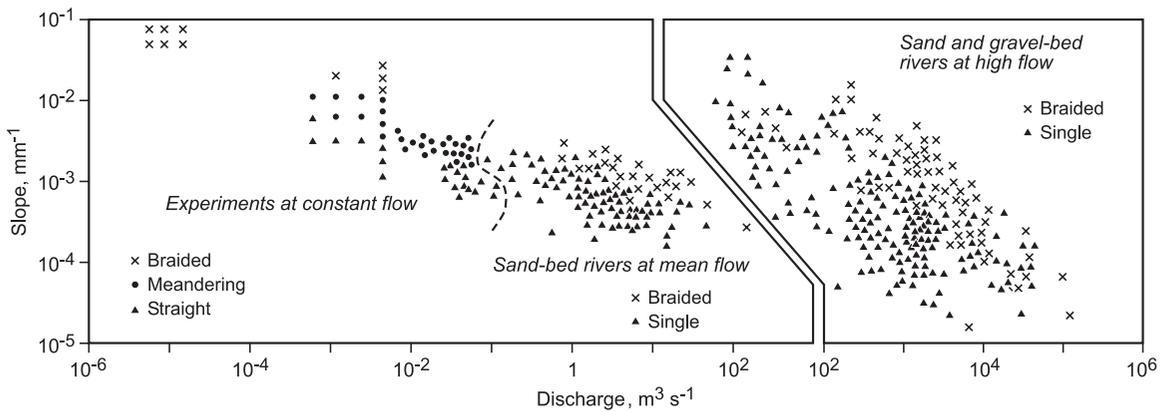


Fig. 12.25 Compilation of field and laboratory data showing the dependency of channel pattern on slope and discharge (after Ferguson 1987). Note that thresholds for meandering in various lab-

oratory models match well and a parallel, higher threshold for braiding is consistent with the experimental results and the data from sand-bed rivers at mean discharge

bedload transport and erodible banks) suggests that it represents a valid equilibrium form.

Field and laboratory data suggest that channel pattern is controlled by discharge and slope (Leopold and Wolman 1957; Schumm and Khan 1972). The compilation of data shown in Fig. 12.25 indicates firstly, that a channel with a given discharge and bed material has one threshold slope (S ; m^{-1}) above which it will meander and a higher threshold slope above which it will braid and secondly, that the threshold slopes decrease as discharge increases. These inverse slope-discharge thresholds have been widely interpreted as thresholds of specific stream power (ω ; $W m^{-2}$) defined as the time rate of potential energy expenditure per unit bed area:

$$\omega = \rho_f g Q S / w \quad (12.14)$$

Ferguson (1981) and Carson (1984) substituted for w using the downstream hydraulic relation $w = aQ^{0.5}$ (Equation 12.11) to show that the meandering-braided threshold corresponds to a constant stream power of $30\text{--}50 W m^{-2}$ (the exact value depends on the value of a). Begin (1981) similarly demonstrated that the slope-discharge threshold also represents a constant shear stress.

The dependency of channel pattern on stream power provides an explanation for the common occurrence of braided rivers in dryland environments. Although flood events in dryland environments are relatively infrequent, they often generate high stream powers as

a result of steep slopes and high discharges. In terms of slope, many dryland streams have high gradients because they flow down pediments and alluvial fans which are steep in comparison to the gradients of valley- and basin-floors. In terms of discharge, high magnitudes are favoured by the effectiveness and efficiency by which rainfall is converted to runoff and concentrated into channels in dryland environments (Baker 1977; Osterkamp and Friedman 2000). In fact, the magnitudes of infrequent flood events in dryland rivers are often much greater than those found in rivers draining humid-temperate catchments of similar size (Costa 1987). Beard (1975), for example, used a flood potential index to demonstrate that the dryland regions of the southwest USA are more susceptible to high magnitude flood events than more humid central and eastern regions. A similar conclusion was reached by Crippen and Bue (1977) who compiled envelope curves for potential maximum flood flows for 17 flood regions of the coterminous USA though their generalisation does not hold for medium-large basins ($> 2,600 km^2$) in which the larger rainfall amounts of more humid regions generate larger floods (Graf 1988, p.90). The tendency for small-medium sized basins in arid and semi-arid environments to have larger floods than similar sized basins in more humid environments is evident in a comparison of flood frequency curves from different climatic regions (Baker 1977; Farquharson et al. 1992; Fig. 12.26).

Channel pattern also reflects sedimentary controls. For example, Schumm (1963) found that channel sinu-

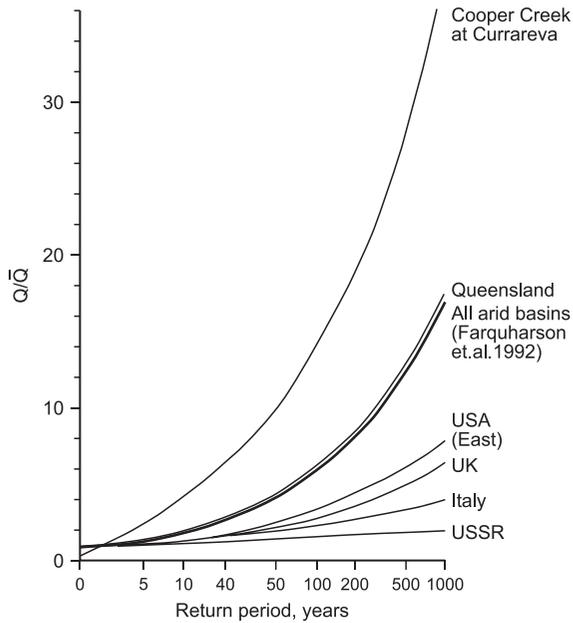


Fig. 12.26 Compilation of flood-frequency relations for rivers from diverse climates showing the more variable flood regime of arid and semi-arid rivers compared to their humid counterparts (after Knighton and Nanson 1997). Note that the magnitude of the mean annual flood (\bar{Q}) event in semi-arid environments is only a few percent of the rare event whereas mean annual events in humid temperate environments are much closer to the rare event. (The mean annual event has a recurrence interval of 2.33 years if the distribution conforms to the EV1 Gumbel distribution)

osity of sand-bed rivers in the Great Plains of the USA increased with the percentage silt clay in the bed and banks and Kellerhalls (1982) and Carson (1984) found that sand-bed rivers braid at lower slopes than gravel-bed rivers with similar discharges. In these studies, grain size is being used as a surrogate for bank strength (as in studies of cross-sectional adjustment discussed earlier) with the implication that channel pattern is dependent on the erodibility of the channel banks as well as the erosivity of the flow (Fig. 12.27).

As already noted, dryland sediments are not generally rich in cohesive silts and clays and unless riparian vegetation is sufficient to stabilise the channel banks and/or discourage the formation of new channels, channel widening will promote the development of a braided channel pattern through the instability of sediment transport in wide channels (Parker 1976). Indeed, Murray and Paola (1994) regard the braided channel form as the inevitable consequence of unconstrained flow over a non-cohesive bed. As such,

vegetation is a primary determinant of channel pattern in dryland environments. The importance of vegetation in controlling dryland channel patterns can be illustrated by reference to the introduction and spread of Tamarisk throughout the SW USA. Tamarisk (commonly known as Saltcedar) was introduced into the SW USA from the Mediterranean basin in the mid 1800s. Because of its competitive advantages over native riparian species (Brotherson and Field 1987; Howe and Knopf 1991), Tamarisk colonised and spread rapidly along riparian corridors and by the late 1990s, had become established in nearly every semi-arid drainage basin within the SW USA (Randall and Marinelli 1996). The result was a marked decrease in channel geometry throughout the SW USA typified by a 27% reduction in the average width of major streams of the Colorado plateau (Graf 1978). Moreover, in rivers widened and destabilised by large flood events, vegetation regrowth and encroachment (including the invasion of tamarisk) resulted in channel narrowing, floodplain reconstruction and the conversion of multi-thread rivers to single-thread forms (Schumm and Lichty 1963; Burkham 1972; Eschner et al. 1983; Martin and Johnson 1987; Friedman et al. 1996; VanLooy and Martin 2005). Although the mechanisms underlying these channel adjustments have not been demonstrated directly in the field, it is generally reasoned that vegetation first colonises and stabilises the surfaces of flood deposits such as channel bars and dunes and which then grow by vertical and lateral accretion of sediment during low magnitude events. Rates of accretion are enhanced by the increased roughness of the vegetated surfaces which reduce flow velocities and encourage sedimentation. In the absence of destructive high flows, these areas of incipient flood plain grow and coalesce, leading to the aggradation and abandonment of the surrounding channels. Over time, a new floodplain develops that is composed of a mosaic of coalesced islands, abandoned channels and areas of floodplain that build up adjacent to the low water channel.

This model of channel evolution is supported by laboratory (Tal and Paola 2007) and cellular (Tal et al. 2003) models of braided rivers which demonstrate how channels choked by vegetation and/or vegetation-induced sedimentation cause reductions in channel width, braiding index and channel mobility (Fig. 12.28). Ultimately, vegetation eliminates weak flow paths, thereby concentrating water into a single

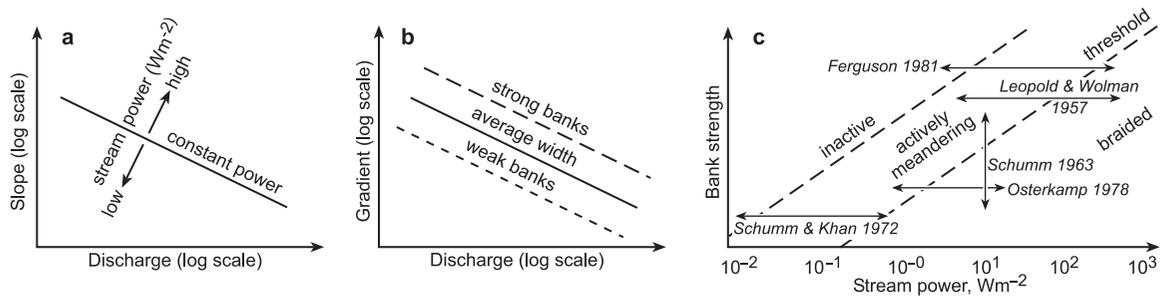


Fig. 12.27 Influence on bank strength on the thresholds of channel pattern (after Ferguson 1987). (a) Threshold of constant stream power (traditional slope-discharge threshold). (b) Anticipated shift in threshold according to bank strength. (c) Empirical relationships between channel pattern, stream power and bank strength. Schumm and Khan (1972) = unvegetated sandy lab-

oratory channels; Osterkamp (1978) = sand bed channels with some silt and clay and riparian vegetation; Schumm (1963) channels with variable silt and clay; Leopold and Wolman (1957) mixture of sand- and gravel-bed channels; Ferguson (1981) = well vegetated banks of gravel and/or cohesive sediments

dominant channel. Progressive reductions in total channel width leads to the establishment of a new self-organised steady state in which the flow removes vegetated areas as fast as they are produced. The results also suggest that colonisation by vegetation is not easily reversible so that the morphological effects are likely to be long-lived. This concurs with the findings of several field studies conducted in dryland environments that suggest that vegetation encroachment raises the threshold for channel adjustment with the result that subsequent floods are not able to widen the channel as they might have previously (Eschner et al. 1983; Hooke 1996). The decreased ability of the channels to adjust to large flood events may lead to increases in the magnitude and frequency of overbank flooding and floodplain sedimentation resulting in further increases in the stability of the channel form through positive feedback (Graf 1978).

Braiding can also be understood as a response to maintain transport competence in relation to the imposed grain size (e.g. Henderson 1961; Carson 1984) and/or transport capacity in relation to the imposed sediment load (e.g. Kirkby 1977; Bettess and White 1983; Chang 1985). From these perspectives, the common occurrence of braided rivers in dryland environments reflects the availability of large amounts of coarse sediment and its movement as bedload. Such approaches provide important links with the underlying causes of braiding, namely local aggradation (often linked to the stalling of bedload sheets, channel bars or loss of competence in flow expansions), bar growth (by vertical and lateral accretion) and subsequent dissection (Ashmore 1991).

Anabranching channels differ from braided channels in that the system of multiple channels is separated by vegetated or otherwise stable islands which are emergent at stages up to the bankfull discharge (Nanson and Knighton 1996). They form a diverse range of channel forms that are associated with flood-dominated channel regimes, resistant banks and mechanisms that induce channel avulsion (Makaske 2001). Low energy, fine grained anabranching systems known as anastomosing channels are found in the Lake Eyre basin (also known as the Channel Country) of east-central Australia (Nanson et al. 1988; Gibling et al. 1998) and in the Red Desert of Wyoming (Schumann 1989). The type is well-represented by Cooper Creek in the Lake Eyre Basin which maintains an active belt of anastomosed channels up to 10 km wide for distances of several hundred kilometres. The floodplain is made up of a well integrated primary system of one-four channels supplemented by subsidiary channels that become active at progressively higher discharges. Anastomosis serves to concentrate stream flow and maximise the transport of bed sediments in regions where there is little opportunity to increase channel gradients (Nanson and Huang 1999; Jansen and Nanson 2004). In other parts of central and northern Australia, channel forms are dominated by ridge-forming anabranching rivers (Tooth and Nanson 2000b). These channels are characterised by a low sinuosity belt of subparallel channels separated by narrow, flow-aligned, sandy ridges vegetated by teatrees (*Melaleuca glomerata*). The ridges develop as a result of spatial patterns of erosion and sedimentation induced by the growth of teatrees within the channel.

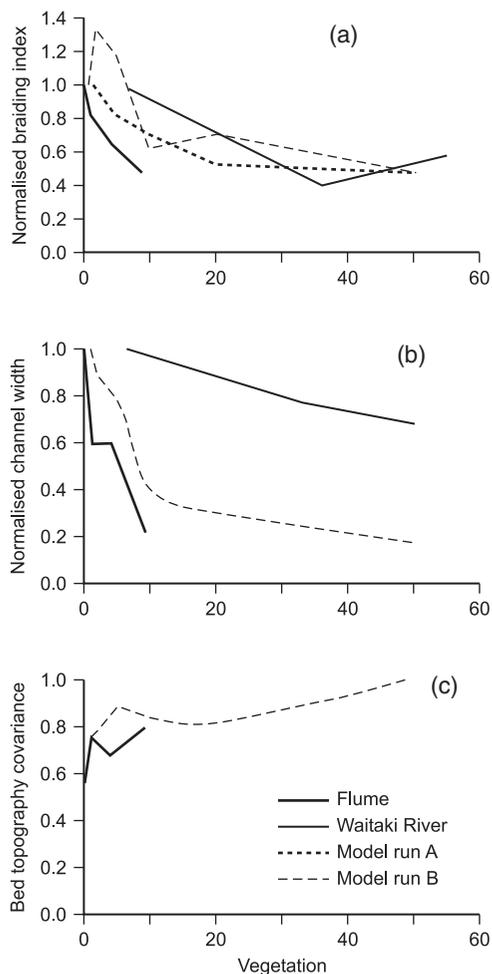


Fig. 12.28 Changes in braiding index (a), channel width (b) and bed topography correlation coefficient (c) with vegetation in the braided Waitaki River, New Zealand, a laboratory flume and a cellular automata model (after Tal et al. 2003). Vegetation in the field, flume and cellular model is represented by fractional vegetation cover of the braid plain, the density of alfalfa stems and a vegetation strength parameter respectively. The bed topography covariance is measure of the channel mobility rate; higher values of covariance indicate lower channel mobility rates

Channel Gradient and the Longitudinal Profile

The longitudinal profile represents the final aspect of channel adjustment to be considered. Longitudinal profiles are typically upwardly concave. It is generally recognised, however, that the long profiles of many dryland rivers are less concave than their humid-temperate counterparts (e.g. Langbein 1964) and may even be linear or upwardly convex (Schumm 1961; Vogel 1989; Fig. 12.29). These differences can be

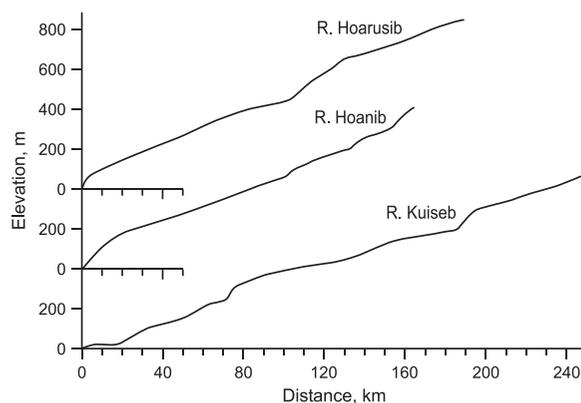


Fig. 12.29 Longitudinal profiles of three rivers in the Namib Desert (after Vogel 1989)

explained by considering the effect of tributary inputs of water and sediment on profile concavity.

In perennial rivers, sediment concentrations usually decrease downstream because tributary inputs increase the supply of water more than they increase the supply of sediment. As a result, stream slopes reduce by incision and an upwardly concave profile tends to develop (Gilbert 1877; Wheeler 1979; Hey and Thorne 1986). In dryland channels, however, the ratio of sediment to streamflow often increases downstream due to transmission losses (e.g. Leopold and Miller 1956). Since sediment is carried by progressively less flow, there is a tendency for aggradation and the development of a convex profile. Profile concavity due to tributary inputs has been modelled by Sinha and Parker (1996). Although Sinha and Parker's model cannot be tested due to a lack of data concerning rates of tributary water and sediment input along river profiles, it does confirm that under conditions of high transport rates, a downstream declining concentration of bed material load is a necessary condition for profile concavity and that convex profiles develop if sediment concentrations increase downstream. The form of the long profile is also a function of sediment calibre and profile concavity has long been associated with the streamwise fining of sediment (Hack 1957; Ikeda 1970; Cherkauer 1972; Parker 1991a,b). However, since downstream fining is the result of particle abrasion and sorting, grain size can control (e.g. Shulits 1941; Yatsu 1955; Sinha and Parker 1996), and be controlled by (e.g. Hoey and Ferguson 1994) profile concavity. The downstream trends of particle size in dryland streams have not been widely studied (notable exceptions include the work of Frostick and Reid 1980; Rhoads 1989). Research in humid-temperate streams has, however,

demonstrated that sediment inputs from tributary inputs and non-alluvial sources often preclude the development of systematic downstream fining trends (e.g. Ferguson et al. 2006). Since high drainage densities and easily erodible banks are characteristic of many dryland rivers, it is quite possible that they may be typified by random grain size variations. Such conditions can be expected to inhibit the development of profile concavity as illustrated by the work of Rice and Church (2001) who studied the profile form of sedimentary links, reaches of river unaffected by inputs of water and sediment. Their results show that under conditions of constant downstream discharge, links with a strong downstream fining trend exhibit concave profile forms (the greater the rate of grain size diminution, the greater the concavity) whilst links exhibiting a weak or no downstream fining trend exhibit convex profile forms.

Summary

As shown by this review, our understanding of dryland rivers has been transformed in recent years through detailed field, laboratory and modelling studies of contemporary runoff and sediment transport processes. However, fundamental questions remain as to how channel morphology and change is related to the time distribution of flow events, variations in sediment supply and their interactions with channel boundary conditions (topography, sedimentology and vegetation) over longer time scales. These questions are not new (e.g. Douglas 1982; Lane and Richards 1997) nor are they restricted to dryland environments (e.g. Kirkby 1999) but they lie at the heart of developing an integrated theory of dryland river behaviour that combines explanations for processes and forms as well as morphological change (Graf 1988). Much work remains to be done to reconcile a kinematic understanding of channel evolution with a dynamic understanding of process mechanics at scales larger than the channel reach and longer than the duration of individual flow events.

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