Rivers and river processes are considered to be one of the most important geomorphic systems on the earth's surface (Leopold et al., 1964; Gregory and Walling, 1973; Schumm, 1977; Morisawa, 1968, 1985; Lewin, 1981). Fluvial systems are among the most dynamic components of landscapes. A number of different (but interacting) variables influence the type and rates of processes operating within a fluvial system (Fig. 3.1.), the single most important being discharge, influencing erosion and transport of particulate and solute load. Water is dynamic, apt to undergo rapid changes of state, and has a high dynamic response rate. Coupled with this, the dynamic characteristics of river systems tend to change in relation to the integral effects of upstream environmental controls. As a result, flow characteristics may vary at any point in the fluvial system. The dynamic nature of river systems is also manifest in dramatic changes in channel morphology and river behaviour, which may result from changes in base level, or from variations in discharge (cf. Schumm, 1977; Chorley et al., 1984). These types of changes, or adjustments, reflect a response of the fluvial system to changes in one or more variables. Response to change may not be immediate. A number of attempts have been made to model catchment response, and to relate the rate of change within fluvial systems to particular variables (Bull, 1975; Yang, 1976; Kirkby, 1977; Knighton, 1984). There are clearly many problems associated with this approach, as a number of the variables are difficult to quantify (e.g., mean climate). For example, in hyper-arid environments, the influence of fluvial processes may be considerable (Schick, 1979), with discharge often resulting from high magnitude, low frequency events (Wolman and Miller, 1960; Baker, 1977; Baker et al., 1983).

In southern Africa (Fig. 3.2.), early research on fluvial systems was concerned with drainage evolution and river capture (Rogers, 1903; du Toit, 1910, 1939; Wellington, 1924, 1938, 1945; Fair, 1944; Dixey, 1945; Maske, 1957; Lenz, 1957) and superimposition (Rogers, 1903, 1925; Wellington, 1926, 1937, 1941; King, 1963; Geyser, 1948). More recent studies have tended to focus on specific aspects and controls of drainage basins, frequently utilising quantitative techniques. Examples of such studies are found in the work on the erosional history of rivers (de Roux, 1968; Tromp and Fuggle, 1969; Twidale and van Zyl, 1981; van Zyl, 1982), river development in karst terrain (Cooks, 1968), the influence exerted by lithology and structure (de Villiers, 1975, 1987; van Wyk and van Rensburg, 1976; Cooks, 1979, 1981, 1983; Garbharran, 1983), and on river morphometry (Russell, 1976; Beckedahl and Moon, 1980). Applied studies have generally been hydrologically orientated (Schulze, 1977), with emphasis
on agrohydrology (Whitmore, 1961), relationships between precipitation and runoff (Abbott, 1976; Pitman, 1976), interception studies (Preston-White and Whittington, 1972; de Villiers, 1976), evapotranspiration (Schulze, 1975), and rainfall variations (Doornkamp and Tyson, 1973). Little attention has, however, been devoted to examining fluvial systems in general (Schulze, 1977; Shaw, 1987).

The aim of this chapter is to examine briefly the present-day hydrological regimes of southern Africa; outline the evolutionary path of some of the major rivers; examine influences of lithology, structure, tectonism and climatic change on the development of southern African fluvial landscapes, and describe the nature of present-day fluvial systems. Finally, adjustments to fluvial systems and their effects on the utilisation of natural resources are assessed.

HYDROLOGICAL REGIMES IN SOUTHERN AFRICA

Discharge within a particular river system is influenced by a number of factors (Fig. 3.3). Of these factors, the timing and distribution of rainfall has a considerable influence on the rate of discharge. With increasing catchment size, however, the output hydrograph becomes increasingly dependent on the integration of responses of tributaries within the catchment, and the relationship becomes less clear.

The flow regime of a river will determine its ability to erode and to transport sediment, and the route taken by water, from input of precipitation to river flow, will influence the solute load. The frequency and intensity of rainfall, and the rate of weathering and erosion processes occurring within the catchment are of particular importance in assessing the geomorphic response of fluvial systems in southern Africa.

Mean annual precipitation in southern Africa is c. 500 mm (compared to a world average of over 850 mm) and mean annual potential evaporation varies between 1 100 and 3 000 mm (South African Weather Bureau, 1986). Less than 9 percent of this precipitation reaches rivers or groundwater storage, and over 90 percent is lost by evapotranspiration (Braune, 1986). Spatial variation in rainfall distribution therefore exerts a considerable influence on hydrological regimes in southern Africa. The intensity and duration characteristics of individual rainfall events (in particular storm-generated precipitation) have a marked effect on stream flow response in southern African rivers (Koace, 1980; Adamson, 1981). This leads to considerable spatial variation in water budgets in southern African rivers. Natal, for example, which occupies 7 percent of the land area of South Africa, accounts for 25 percent of total average streamflow.

The entrainment and transport of sediment by rivers is most dramatic during peak discharge because of the high energy conditions and improved hydraulic efficiency (the later allowing the channel to accommodate the increased discharge). In terms of gross sediment yield, however, the availability of transportable material appears to be more important than fluvial transporting capacity (Roseboom, 1984; van Wyk, 1986; Moore, 1987). Between 100 and 150 million tonnes of sediment are discharged annually by the rivers of southern Africa south of the Limpopo (Moore, 1987). These high sediment yields have caused considerable problems in the loss of reservoir storage capacity.

EVOLUTION OF FLUVIAL SYSTEMS IN SOUTHERN AFRICA

Fluvial systems evolve over time under a variety of environmental conditions. Their stage of evolution depends on a number of variables (Fig. 3.4). These influence the progress of the denudation of a landscape and its hydrology (Schumm and Lichty, 1965; Schumm, 1972, 1977; Starkel and Thorns, 1981). One of these variables, time, has considerable significance in the southern African landscape. The sedimentary record indicates that the landscape has been under predominantly fluvial influence without major interruption since the early Mesozoic (Tankard et al., 1982), and subject

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**Fig. 3.3:** Major controls of discharge characteristics in a fluvial system (after Jones, 1983).

**Fig. 3.4:** An idealised model of a fluvial system, showing major drainage system variables (modified after Schumm, 1977 and Schumm and Lichty, 1968).
to denudation primarily by fluvial processes since the early Triassic (Rogers, 1903). A number of the larger rivers (e.g., Orange, Limpopo) are thought to have been evolving since the Triassic, though pre-Karoo drainage lines have been widely identified (du Toit, 1910; Rogers, 1934; Wellington, 1924, 1929, 1933, 1955, 1958; Partridge and Brink, 1967; Petroc, 1976; Helgren, 1977, 1979; McCarthy, 1983; Marshall, 1987).

Since the Triassic, southern Africa has been subjected to large-scale changes of topography (cf. Dixey, 1945; King, 1963, 1972), coincident with the break-up of Gondwanaland. Changes in vegetation type since the late Cretaceous may have resulted in significant changes in runoff (cf. Gevers, 1948; Wicht and Banks, 1963; van Wyk, 1987) and sediment yield (Langbein and Schumm, 1958). It is likely that rates of denudation have varied considerably in space and time, dependent on changes in relative relief associated with base level changes (cf. Baillie, 1969, 1970; Watson and Price-Williams, 1985), climate, geology, vegetation, hydrology and sediment yield. It should be borne in mind, however, that evidence concerning the palaeohydrology of the remote past is nebulous, and often biased to the depositional rather than to the erosional history of a fluvial system, so that it may have little value in shedding light on the overall nature of post-hydrosic regimes (Schumm, 1968; Hickin, 1983).

Pre-Cainozoic fluvial systems in southern Africa formed under hydrological conditions different from the present. It is likely, given the very different pre-Cainozoic botanical (i.e., the absence of angiosperms) and climatic framework in southern Africa (cf. Coetzee et al., 1983; Tyson, 1986), that rates of erosion and sediment yield were significantly different from the present day (cf. Noble, 1965), in some instances as much as a hundredfold greater (Storey et al., 1964). Much of the sedimentary evidence from the Permo-Triassic Karoo basin (Hobday, 1978; Hobday et al., 1975; Smith, 1978; Turner, 1980; Visser and Botha, 1980; Stear, 1983) shows the predominance of low-sinuosity channels, levee, interchannel and crevasse splay sub-environments, associated with extensive alluvial plain environments. This suggests predominance of bedload deposition coupled with high sediment yields, possibly associated with limited primitive vegetation cover (Schumm, 1968).

It is possible that the morpho-dynamic response of arid-region hydrological regimes (cf. Hooke, 1967) may be determined principally by reduced vegetation cover. In present-day environments reduced vegetation cover is often associated with either aridity or human interference. In pre-Cainozoic times this may not have been the case. It is likely, therefore, that pre-Cainozoic palaeohydrology in southern Africa was determined largely by sparsely distributed biota, producing high peak discharges and high sediment yields, resulting in characteristic braided or anastomosing channel forms developed in predominantly alluvial plain environments.

Similar fluvial depositional environments exist at present in unvegetated hyper-arid environments (cf. Schick, 1979), and are characterised by widespread development of alluvial fan and playa lakes (Fig. 3.5). These often form catastrophically, being associated with high-intensity storms of short duration (Gerson, 1982). The 6 200 m² alluvial fan of Wadi Mikeimin, in eastern Sinai, for example, formed overnight (Lekach, 1974). Work on alluvial fan deposits in southern Africa (Booyzen, 1974; Odendaal, 1979, 1980) suggests that large quantities of alluvium were deposited rapidly, probably during wetter phases of the Pleistocene (Booyzen, 1974). A detailed review of arid-zone alluvial fans is given in Chapter 5.

The break-up of Gondwanaland and related tectonic flexuring produced major changes in the fluvial systems of southern Africa. Post-Mesozoic estrangement from major source areas for debris south of the present coastline resulted in a change from predominantly fluvial depositional to fluvial denudational landforms (Helgren, 1979b) (Fig. 3.6).

Environmental reconstruction of post-Mesozoic fluvial drainage networks has been fragmentary in southern Africa (Rogers, 1903, 1925; Wellington, 1955; van Zyl, 1982). Most reconstructions have been based on the evaluation of erosion scars, abandoned channel remnants, erosional and depositional terrace remnants and evidence of changing patterns of nearshore marine sedimentation (Tankard, 1976; Helgren, 1979b; Twidale and van Zyl, 1981; McCarthy, 1983; Dingle and Hendey, 1984).
Present-day fluvial systems in southern Africa are as varied and complex as their Cainozoic and pre-Cainozoic counterparts. Almost every present-day fluvial network in southern Africa has inherited features or controls, so that it is impossible to isolate predominant influences within particular time periods. They appear to be influenced by factors such as topography, vegetation, land-use and climate-hydrologic regime. Some controversy exists regarding the influence of geological control on present-day systems (cf. Cooks, 1979, 1981, 1983; van Wyk and van Rensburg, 1976). Human interference is also having increasing influence on fluvial systems (Morgan, 1971; Gregory, 1974; Barnard, 1978).

It is pertinent, therefore, to consider fluvial networks in terms of long-, medium- and short-term developments, rather than to isolate particular elements or periods of development.

Drainage systems of the eastern escarpment
The eastern escarpment is drained mainly by east-flowing rivers, the exception being the Limpopo-Olifants system, which also drains much of the northeastern and central Transvaal. High relief ratios on the eastern escarpment have produced relatively small drainage basins compared with the interior, characterised for the most part by high-gradient, high-sinuosity streams (Fig. 3.7). Drainage lines are for the most part structurally controlled or superimposed. (Relations between structure and drainage are discussed in Chapter II). The drainage systems on the eastern escarpment are the direct consequence of post-Mesozoic tectonism (King, 1972). The drainage networks are at present degradational with considerable down-cutting and limited sediment storage within channels (Fig. 3.8), although this may not have been the case during early Pleistocene times (Bailie, 1969, 1970; Watson and Price-Williams, 1985). Incised meanders predominate, largely independent of topography.

The Limpopo system, in contrast to most other east-flowing systems, is seasonal or even episodic, and largely aggradational in character. Many of the tributaries have sand-choked beds, and although the river valleys are generally wide, the channels themselves are incised by as much as ten metres (Hugo, 1970). The aggradation in parts of the system has variously been ascribed to variations in rainfall, the destruction of natural vegetation in source areas, agricultural and forestry practices, and increases in soil loss due to human factors (Wellington, 1929; Gevers, 1948; Hugo, 1970).

Drainage systems of the interior
Interior drainage systems are very extensive. The Orange River catchment is 953 200 km² in area (Dingle and Hendey, 1984) and is one of the largest drainage basins in the world. The Okavango system is similarly large and forms one of a small number of closed hydrologic systems found in southern Africa (Shaw, 1987). These systems are generally characterised by low gradients, though incised meanders occur within parts of the systems.
Studies of sediments within the Orange Basin indicate that the upper Orange-Vaal system entered the south Atlantic through the Cape Canyon, some 300 km south of its present mouth, during the Oligocene (Dingle and Hendey, 1984) (Fig. 3.9b). This is supported by terrestrial geomorphic evidence (McCarthy et al., 1985). The present lower Orange River drained a relatively small area (Fig. 3.9b), and the upper Orange flowed further south as part of the present Krom River with the Olifants River (Cape Province) as a tributary. Successive river capture by the Koa River and its tributary headwaters resulted in the diversion of the upper Orange to its present course (Fig. 3.9c) (Dingle et al., 1983; Hendey, 1983). The river capture and subsequent abandonment of the flow through the Koa valley is thought to be associated with tectonic warping (du Toit, 1933), which affected drainage elsewhere on the continent (Mayer, 1973; Stratten, 1979; McCarthy, 1983), and was probably enhanced by intrusive volcanism during that time (Moore, 1979).

Mean annual sediment yield for the Orange River system shows a progressive decrease from $10 \times 10^7$ m$^3$ in the late Cretaceous, to $3.86 \times 10^7$ m$^3$ in the Palaeogene, to $0.3 \times 10^7$ m$^3$ in Neogene times (Dingle and Hendey, 1984). Sediment yield in the Neogene period has been interpreted as reflecting desert conditions (Milliman and Meade, 1983) and, although weak sporadic upwelling of the proto-Benguela Current during the Oligocene has been correlated with increasing aridity (with precipitation levels intermediate between Cretaceous and Neogene times) (Siesser, 1978), the decrease in rates of sedimentation could also be explained by the abandonment of the Trans-Tswana River (McCarthy, 1983). This river is thought to have

The present Orange River system has a mean annual discharge of $1.061 \times 10^7$ m$^3$, and has been in existence since the Cretaceous. Analysis of sediments in the southern Atlantic suggests that the Orange Basin was the main depocentre for Mesozoic and Tertiary sedimentation, having formed during continental break-up in early Cretaceous times (Dingle and Hendey, 1984). The geomorphic development of the region drained by the Orange River system has been the subject of several studies (cf. Wellington, 1933, 1958; Geyser, 1948; Mabbutt, 1955; King, 1963; van Rooyen and Burger, 1973). That there have been major course changes during the river's history was first postulated by du Toit (1910). The suggestion of a more extensive river system than the present has been cited more recently to explain similarities in modern freshwater fish fauna

![Fig. 3.8: High mountain stream gullies, showing relatively little sediment storage or stream deposition, Natal Drakensberg.](image)

![Fig. 3.9: Reconstruction of palaeodrainage lines of the Orange River drainage systems, (a) late Cretaceous, (b) Palaeogene, (c) Neogene (simplified after Dingle and Hendey, 1984).](image)
entered the Orange River system from the north, in the general vicinity of the present Sand River (southwest of Douglas).

Evidence for a large river, possibly with as much as four times the discharge of the Orange River, is based on provenance studies of river gravel and terrace deposits. The sediment volume discharged by the Trans-Tswana River suggests a catchment area that must have drained much of south-central Africa (McCarthy, 1983). Crustal warping (Transvaal-Griqualand axis, Fig. 3.10) and northerly sinking (cf. du Toit, 1933) extends across the inferred course of the Trans-Tswana River. If warping had been rapid, it might have led to the formation of an internal drainage system (such as the Okavango River). McCarthy (1983) suggests that the demise of this system was associated with infilling of the Kalahari basin. It has also been suggested that the Okavango Delta, Lake Ngami and the Makgadikgadi Pans are the final remnants of this once vast drainage system (McCarthy, 1983) (Fig. 3.11). Early workers have hypothesised a link between the Okavango and Limpopo or Zambezi Rivers during mid-Tertiary times (Rogers, 1936; van Straten 1963; Cooke, 1976; Wright, 1978). Wellington (1955) and Grove (1969), however, suggest that the Upper Zambezi and Lupanga Rivers flowed southwards, contributing sediment to the Kalahari basin prior to being truncated by successive capture by a proto-lower Zambezi.

![Fig. 3.10: Crustal warping associated with the Trans-Tswana River system (modified after Mayer, 1973 and du Toit, 1910).](image)

Evidence suggests that diastrophism along the Griqualand-Transvaal Axis during the Tertiary is responsible for the occurrence of high-level gravels along the Vaal River system (Mayer, 1973), rather than higher discharges resulting from climatic variability during the Pleistocene (Sohng et al., 1937; Cooke, 1946; du Toit, 1954; Mason, 1967; Haughton, 1969). The latter may, however, have accelerated the derangement and abandonment of drainage lines. The Vaal and Harts Rivers of the Orange River system have been extensively studied in consequence of their association with alluvial diamonds (cf. Johnson and Young, 1906; du Toit, 1900, 1910, 1951; Haughton, 1921; Sohng et al., 1937; Cooke, 1946; Helgren, 1977). The stratigraphy and genesis of alluvial terraces in the region has been studied by Partridge and Brink (1967) and Helgren (1979), whereas the morphometric characteristics of the diamondiferous gravel have been studied by Stratten (1979) and Holmes (1987).

The upper reaches of the present Harts River and several of its tributaries may have been heaved by the upper reaches of the Molopo River due to rapid headward erosion related to tectonism (Mayer, 1973).

Extensive pan sequences occur between the Molopo River and the Makgadikgadi Basin and, although at present related to deflation (cf. Grove, 1969; le Roux, 1978; Lancaster, 1978) their origin has been ascribed to fossil drainage lines (Geyser, 1950; Boocock and van Straten, 1962).

The Okavango Delta is a fault-controlled alluvial fan of 22 000 km² in northwestern Botswana. It is subject to seasonal flooding from 6 000 to 13 000 km² (UNDP/FAO, 1977). It has an annual recharge of 15.5 × 10⁶ m³ yr⁻¹, with an annual sediment input of 400 000 m³ (McCarthy et al., 1986a). Flow patterns within the Delta are erratic, varying in response to changing patterns of sedimentation, minor

![Fig. 3.11: A summary of hypothetical reconstructions of the Trans-Tswana and related paleodrainage systems (after Wellington, 1955; Grove, 1969; Mayer, 1973; Swaneveldt, 1974; McCarthy, 1983; Dingle and Hendey, 1984, and McCarthy et al., 1985).](image)
tectonism, vegetation blockage and activities of aquatic animals (Shaw, 1987b; McCarthy et al., 1986a, 1986b, 1987; Dincer et al., 1978).

On the basis of strand lines, Grove (1969) suggests that the Makgadikgadi Pans at one stage formed a palaeo-lake with an approximate area of 34 000 km². Lake Ngami was also considerably more extensive in the recent past, with an area of up to 1 040 km² (Shaw, Chapter 7). The Etosha Pan was also more extensive, fed by the Kunene River prior to its capture and diversion through to the Atlantic Ocean (Grove, 1969) (Fig. 3.11).

**Drainage systems in Namibia**

Fluvial systems in Namibia are influenced by the arid and semi-arid conditions that prevail over much of the area. Westward-flowing drainage systems appear to be subject to regional control, possibly related to climatic, tectonic or eustatic changes (Ward, 1982). Wilkinson (Chapter 5) has distinguished two main types of drainage systems (exoreic and endoreic drainage) in this area. Some rivers, such as the Orange River, are clearly of the exoreic type. However, drainage type in this area appears to be influenced directly by patterns of dune migration or by climate, and is therefore liable to change, both in the short and long term. Exoreic drainage lines, for example, have become endoreic (i.e., internal drainage) as a result of dune migration northward during the late Pleistocene (Seely and Sandelowsky, 1974). The Dune Namib has cut off a number of streams in this way (e.g., Tsodab, Tsauachab), which now form internal drainage lines, often terminating in “vleis” (Fig. 3.12). It appears likely that the former drainage lines underlying the Dune Namib may in some cases act as conduits for groundwater seepage to the Atlantic (Swanevelder, 1974).

Drainage in Namibia is largely episodic or ephemeral (Swanevelder, 1974; Marker, 1977; Ward and Breen, 1983) with only the Omaruru, Swakop-Khan and Orange Rivers able to maintain courses to the Atlantic (Marker, 1977), although the Swakop-Khan system does not at present debouch directly into the sea. Even the Orange estuary is continually migrating across a beach barrier (van Heerden, 1986). The Kuiseb-Gaub system is intermediate in character (Marker, 1977).

**ADJUSTMENTS TO FLUVIAL SYSTEMS: CHANNEL FORMS AND CHANNEL CHANGES**

The foregoing discussion indicates that fluvial systems in southern Africa are subject to change in both time and space. On the one hand, the erosional and depositional components of the fluvial system may evolve progressively through long periods of time, with major morphological changes resulting from tectonism or climatic change (Schumm, 1977; Shaw, 1987b). On the other, adjustments that are independent of tectonism or climatic change may also be observed. These occur where threshold values of stress are exceeded (cf. Schumm, 1973), and can result from incipient instability within the fluvial system or where gradual increases in external stress are applied, so that minor additions of stress result in large-scale changes within the fluvial system. Schumm (1973) has demonstrated, for example, that gully development (see Chapter 12) may be initiated where critical slope angles are exceeded within a drainage basin (Fig. 3.13). More catastrophic adjustments have been postulated for changing patterns of gully development (Graf, 1979; Haigh, 1987).

![Fig. 3.12: Drainage systems in Namibia (modified after Marker, 1977).](image)

![Fig. 3.13: Relationship between valley slope and drainage area, Piceance Creek basin, Colorado, showing a critical threshold for gullying (modified after Schumm, 1973).](image)
Channel form can be viewed as the outcome of the interaction between the erosive potential of a river and the resistive forces operating within the drainage basin (Ferguson, 1981). Channel geometry (i.e., river longitudinal profile and slope; channel cross-section and size, and channel patterns) is influenced by a number of geological, hydrological, anthropogenic and inheritance factors.

Structural controls on drainage patterns in southern Africa have been outlined by Moon (Chapter II) and Cooks (1972), and will not be discussed in detail here. Channel patterns will also show short- to long-term adjustments. Schumm and Kahn (1972) have demonstrated, under experimental conditions, that channel form will change at threshold values from straight to meandering to braided, with increasing slope angle. Similar changes in channel form have been noted in the Mpmabeni River in Zululand, resulting from artificial changes in channel gradient from 1:180 to 1:600 (Dix, 1984). Abrupt changes in channel form may also result from major floods as occurred in the Tugela River following Cyclone Domaina, in January 1984 (Kovacs et al., 1985).

Though it is possible to observe short-term adjustments such as gully development or channel pattern changes, in reality little is known about the response of a fluvial system to long-term events. It has been demonstrated, for example, that one or more events can trigger complex reactions as the components of the fluvial system respond progressively to change (Lewis, 1949; Schumm and Parker, 1972). Thus an infrequent but relatively large-scale stress (e.g., a flood or drought) may act as a catalyst to exceed a geomorphic threshold and trigger a sequence of events that will result in significant landscape modification. Frequent small-scale stresses (e.g., over-grazing, footpath erosion) may have a similar effect (Fig. 3.14).

The longitudinal profile of rivers in southern Africa has not been examined in detail, either in terms of compound profiles or with the aim of establishing concavity indices (cf. Wheeler, 1979). King (1972) has, however, pointed out the distinctive changes in the longitudinal profile of rivers in Natal and Transkei, which are most probably associated with changes in eustatic sea levels or tectonism during the Cainozoic. Such changes are also manifest in the abundance of raised terraces and alluvial deposits occurring in the Natal Drakensberg (King, 1972) and in the interior (Helgren, 1979a), and through channel abandonment (Fig. 3.15), which formed as a result of base level changes, or climate-related changes in discharge.

THE EFFECTS OF EXTREME EVENTS ON FLUVIAL SYSTEMS

Fluvial systems are often subject to rapid changes, particularly in terms of discharge and sediment yield. In southern Africa, extreme weather events (e.g., cyclones, drought) are often closely reflected in the responses of fluvial systems. The Pretoria flood in 1978, for example, which resulted from 280 mm of rainfall occurring within an eight-hour period, had a short concentration time (due to the high degree of urbanisation in the flood region), produced a flood peak flow rate of 10.7 km hr⁻¹ (Kovacs, 1978).

The Laingsburg flood (January 1981) resulted from soaking rainfall lasting three days with localised short-duration, high-intensity rainfall. There were very high discharges, with 90 per cent runoff occurring towards the end of the rainfall. Flood peaks of 1 450 m³ sec⁻¹ and 5 680 m³ sec⁻¹ were recorded at Robertson and Laingsberg respectively (Kovacs, 1982). The high discharges resulted in considerable erosion and deposition. At peak discharge, 5 750 m³ sec⁻¹ of water entered the Floriskraal Dam below Laingsburg. Seventy per cent of the sediment carried by the river (i.e., 13.2 million m³) remained in the dam after the flood event. The Domaina floods (January/February 1984), resulted from rainfall spread over a five-day period. The three-day 200 year rainfall event was exceeded over an area of 32 000 km², with more than 400 mm of rain falling over an area greater than 47 000 km² (Kovacs et al., 1985; van Heerden and Swart, 1986). Flood peaks of 16 000 m³ sec⁻¹ from a catchment of 9 126 km² were recorded on the Mfolozi River. Runoff was in the order of 70 per cent of rainfall, and flood wave propagation velocities reached 25 km hr⁻¹. Sediment yields were extremely high, with c. 100 million m³ of sediment being deposited in the Pongolapoort Dam. During the flooding of the Mfolozi,
brought about by the application of infrequent large-scale stresses or frequent small-scale stresses, or both. Human influence on fluvial systems may be direct (e.g., reservoirs, river diversions, channel straightening, bank protection, catchment transfers) or indirect (e.g., land use, urbanisation).

Reservoirs have considerable influences on fluvial systems, through reductions in the magnitude and frequency of runoff, through increased evaporation, through restricted sediment transport, and through increased scour downstream. Inter- and inter-basin transfers of water (e.g., Orange-Fish-Sundays and Tugela-Vaal schemes) represent major changes in natural fluvial systems. The Orange-Fish-Sundays transfer scheme (Fig. 3.16) has decreased the net flow of the Orange River and increased the flow of the Fish and Sundays Rivers (Tylcoat and Forster, 1987). Indirect pumping of water from rivers or from aquifers near to rivers reduces flow volume. This can inhibit base flow in small streams, resulting in net reduction in discharge within

**Fig. 3.15:** Cut-off meander resulting from down-cutting, Mzimvubu River, Transkei.

a sediment wedge of 1-2 m (characterised by shallow-water deltaic sedimentary facies) was deposited on the coastal flat lands. The estuary of St. Lucia was subject to considerable scouring (to a depth of 6 m) (van Heerden and Swart, 1986).

The impact of drought on fluvial systems is more difficult to establish. Drought periods in southern Africa are characterised by reduced geomorphological activity of fluvial systems, but may provide greater opportunity for some forms of weathering to take place, thus increasing the amount of material susceptible to fluvial erosion and transport (Zucchini and Adamson, 1984).

**MANAGEMENT OF FLUVIAL SYSTEMS**

The dynamic aspects of fluvial systems have been a theme of this chapter. It has been suggested that most fluvial systems are in dynamic equilibrium, subject to changes

**Fig. 3.16:** The Orange-Fish-Sundays River transfer scheme (after Tylcoat and Forster, 1987).
the catchment. Channel modifications in rivers may increase hydraulic efficiency. The effects may be beneficial at the point of modification but the consequences further downstream could be increased deposition or increased flood risk.

Indirect human influences are more difficult to assess, as changes are more gradual and geomorphological adjustments may take considerable time. Changes in land use can cause marked changes in runoff by, for example, reduction in infiltration rate resulting from degradation of vegetation cover (e.g., overgrazing, burning). Afforestation may, in contrast, result in a reduction of flood peaks and an increase in recession flow (van der Zyl and Walker, 1986).

Urbanisation effects on flow quality and quantity are important because of the critical water needs of southern Africa. Most attention in southern Africa has focused on the high costs of treating water affected by urban environments (Pitman and Watson, 1984), and on the problems of high-peak runoff resulting from the increased area of paved surfaces and the dense artificial drainage networks found in urban areas (Beckedahl, 1981; Stephenson et al., 1986).

**WATER RESOURCES**

Because southern Africa is water deficient, with a long history of support problems, considerable attention is devoted by hydrologists, engineers and planners to water resource problems (cf. Maaren, 1984; Hughes and Stone, 1987). Much attention has been focused on engineering projects to regulate water supply, and to maximise the use of river discharge. Figure 3.17 illustrates the water resources problem facing South Africa. Relatively few new sites suitable for dam construction remain, and existing dams are being infilled with sediment. A constant rise in the demand for water is aggravating the problem.

River systems are now recognised as important ecological entities (O’Keeffe, 1986), within which total catchment management is essential both to conserve and to develop resource potential (van der Zyl and Walker, 1986). There is a considerable investment in instrumentation for measuring data concerning the interrelationships between hydrological variables in southern Africa. As techniques become more refined, the assessment of the water resources potential of southern African rivers may become even more accurate (Midgley et al., 1983). Solutions to the demand-supply problem will require greater and more efficient use of groundwater, more efficient use of water by agriculture and industry, and water management coordinated at a subcontinental level.

The influence of humans on river systems in southern Africa is enormous (and evidently increasing), and in all but the smallest of rivers the fluvial system has been altered by land-use modifications or by engineering structures. Extreme events, such as the floods associated with Cyclone Domoina, serve as a reminder, however, that the natural forces shaping the earth are, in the final analysis, much greater than those of the human race.

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4 Hillslope Form and Development

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For many decades the study of slopes has been a central theme in geomorphology and the origins, development and forms of hillslopes pose some of the fundamental problems in the discipline. Since landforms and landscapes are combinations of slopes of a variety of forms it follows that an understanding of slope form and development is essential in geomorphology. The importance of slope geomorphology as a sub-discipline is further reflected in the extensive literature on the subject (for example, Young, 1972; Carson and Kirkby, 1972; Small and Clark, 1982; Schly, 1982a). Themes such as a slope form, slope evolution, the operation and effects of slope processes and the relationships between forms and processes, and rock and regolith properties as controls of slope development have received considerable attention.

In this chapter attention is focused on hillslope form and development in the southern African landscape. Initially an outline of the theories of slope development is presented and conditions promoting different types of slope development are considered. To provide a link between slope theory and local landscapes, slope development theory is reviewed in relation to examples of slope evolution under southern African conditions.

In order to understand slope development theory it is necessary to have an appreciation of the different forms that slopes may assume, and of the nature of the processes inducing changes of those forms. Slopes may vary from simple, single-element forms (either rectilinear, convex or concave) to complex profiles composed of different elements (Fig. 4.1). Slopes may form on bare rock, but more usually develop on weathered and transported material.

The processes contributing to changes of slope form are rock weathering (chemical, mechanical and biotic) and erosion induced by water, wind, ice and gravity. The erosional processes are divisible into the gradually and rapidly operating types. Gradual processes comprise rainsplash erosion, creep and wash processes, and solution; and the more rapid processes involve falls, slides and flows of material at different scales (Sharpe, 1938). The rates at which surface processes operate vary according to environmental conditions (in that they determine the type and intensity of process), the nature of the material of which the slope is formed, the relief, and the gradient of the slope. Process rates have come under close scrutiny over the last few decades. The extent of this concern with process rates is reflected in the collation and review of the results of some 400 such studies (Saunders and Young, 1983; Young
Arid Landscapes

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Geomorphologists have long been impressed by the fact that climate has exerted strong control on landform development. In the case of glacial and arid landscapes the climatic control cuts across other controls such as rock type, rock structure and tectonic history. Those features that are particularly closely allied to aridity (especially under warmer climates, since cold deserts are absent in southern Africa) are described in this chapter.

The chapter defines deserts for the purpose of geomorphology. Three major topics are considered, those of water-moulded landscapes, wind-sculpted landscapes, and processes operating on comparatively inactive surfaces, including two soils common to southern African landscapes.

DESSERT CLIMATE, VEGETATION COVER AND SEDIMENT YIELD

Because of the importance of temperature in controlling the effectiveness of rainfall (rain in hot deserts can often be seen evaporating long before it reaches the ground), recent climatic classifications have incorporated temperature in defining boundaries between areas of one kind of moisture effectiveness and another. In deserts, potential evaporation can be two or even three orders of magnitude greater than rainfall. One widely used classification system is that of Meigs (1953) whose rainfall-evaporation ratios are used here. By this system extreme aridity occupies 4 per cent of the world's surface; 15 per cent is arid and 15 per cent semi-arid (Cooke and Warren, 1973). In southern Africa this means that those areas to the west and south of the 300 mm isohyet are semi-arid or drier, the area of extreme aridity encompassing places where no rainfall has been recorded for at least 12 consecutive months (Fig. 5.1). Rainfall increases with distance from these desert cores and becomes more seasonally rhythmic. The semi-arid areas (in southern Africa's particular temperature regime) lie approximately between the 300 and 500 mm isohyets. One author has classed geomorphic aridity as those areas with "ephemeral" or "sporadic" stream flow, and semi-arid areas as those with regular seasonal flow (Butzer, 1976).

Worldwide patterns of aridity apply equally to southern Africa, the desert areas being confined broadly to the zonal belt of subtropical, semi-permanent high pressure cells. Regional controls modify this pattern and produce an asymmetry with aridity in the western half of the subcontinent. Local controls in turn distort this pattern, so that major river valleys are drier than surrounding higher country (e.g., the Tugela and Limpopo River valleys). Rain shadow areas such as the Karoo immediately in-

Fig. 5.1: Subcontinental zones of extreme aridity, aridity and semi-aridity (after Meigs, 1953).