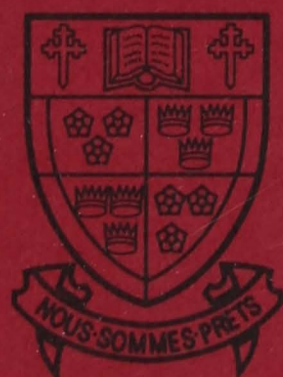
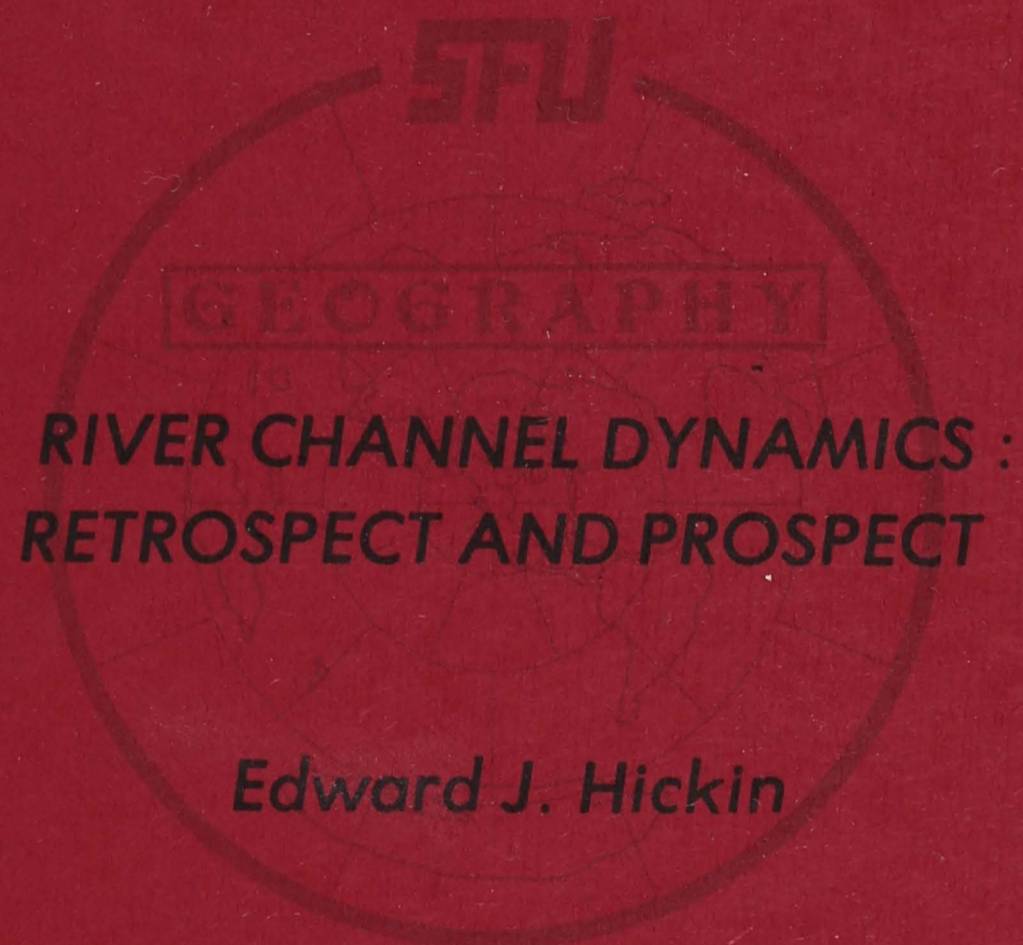


**DEPARTMENT
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DISCUSSION
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River Studies: Part III

River channel dynamics: retrospect and prospect

by

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Comments are invited.

Preface and Acknowledgements

This paper is the final part of a series on the basic principles and theories of river behaviour. Part I (Discussion Paper No. 9) reviewed the topics of drainage basin morphometry and river network analysis and Part II (Discussion Paper No. 10) reviewed elements of drainage basin hydrology.

This paper attempts to define the state of the art in the field of channel dynamics, to identify critical problem areas, and to suggest the directions of future research.

What follows is a personal view of recent developments in fluvial geomorphology. Although I am responsible for its inadequacies, its strengths owe much to the stimulation provided by my students and colleagues of the last 15 years.

Although the manner in which rivers change the form and pattern of their channels in response to environmental change has been a recurring theme in river studies, it recently has enjoyed considerably increased attention from earth scientists. Perhaps the most significant recent evidence of this interest is the appearance of several collected works and reviews of studies of channel changes (for example, see Gregory, 1977; Gregory and Walling, 1979; Kuprianov and Kopaliani, 1979; Park, 1981), and the fact that a Session has been devoted to the topic here at the Second International Conference on Fluvial Sediments at Keele, England, in September, 1981.

The study of river channel changes, in the broadest sense of the term, is no less than the study of equilibrium channel behaviour and the nature of excursions from those equilibrium conditions. As such it includes almost all that we know about the fluid mechanics and morphology of alluvial channels. But in a more narrow sense of the term it is the collection of empirical and theoretical studies concerned with the adjustment of channel cross-sectional size, form, and pattern, to shifts in environmental conditions, particularly those that promote changes in discharge and in sediment loads. In a still narrower sense, channel changes may be regarded as those that have been induced by the activities of human beings.

Channel Equilibrium and Time Scales

Central to any discussion of river changes, however, is the concept of equilibrium, and as Schumm and Lichty reminded physical geologists in 1965, geomorphic equilibrium only has meaning in the context of particular time scales. The full significance of this fact only now is achieving general recognition. Organising knowledge by time scale may be useful and desirable for many purposes but it also constrains the way in which we think about the world. Time scale selection largely determines the questions that we can ask. In the context of channel changes it has emerged as a serious limitation to our understanding of the ways of rivers.

Changes in river form and pattern typically have been examined at three arbitrary time scales; for the sake of discussion we might call them geologic, geomorphic and engineering time scales.

Geologic time, measured in terms of perhaps a million plus years, might also be termed cyclic time; it is the time scale in which the completion of an erosion cycle is a significant event. The governing conditions for landscape change at the geologic time scale include major tectonic events such as crustal plate displacement and related processes of regional warping and mountain building. These factors, in turn, influence climate, weathering, and general denudation rates. It is the time scale that many of us will associate with the regional physiographic models of classical geomorphologists such as William Davis, Lester King, and Walther Penck. It is the domain of the physical geologist or regional physiographer who relies on the rock record, on global radiometric dating, on earth magnetic-field reversals, on surveys of planation surfaces and of deep weathering, and on the methods of structural and historical geology, to piece together the evolution of the planet's surface.

The geomorphic time scale is that of much of modern fluvial geomorphology, ranging from a hundred to a few hundreds of thousands of years. For example, the now classical palaeoclimatic reconstruction by G.H. Dury (1953, 1954, 1955, 1958, 1960, 1964(a), 1964(b), 1965, 1976) largely is based on Holocene and late Pleistocene river underfitness. Similarly, later attempts at refinement of this method by Schumm (1960, 1963, 1965, 1968, 1969) largely are based on studies of contemporary river behaviour on the American Great Plains and on several tens of thousands of years of history of the Murrumbidgee River in the Riverine Plains of Eastern Australia.

Fluvial geomorphologists generally have assumed that the governing conditions for landscape change at the geologic time scale are sensibly constant and therefore largely irrelevant to discussions of river changes over geomorphic time. They have been far more concerned with the fluvial consequences of the profound environmental fluctuations associated with Pleistocene glaciation. Investigations of the history of river changes evidenced by valley fills, terraces, deranged drainage, channel-capacity contraction, and so on, have relied on extensive field observation employing the descriptive tools of the stratigrapher and soil scientist. Much of the interpretation has relied on the use of techniques such as the mechanical analysis of sediments, C^{14} dating, pollen and tree ring analysis and more recently on identifying actual and laboratory analogues through studies of contemporary river form and process. It is the latter development that has seen research interests of geomorphologists, sedimentologists and engineers converge in recent decades.

Fluvial geomorphologists have also relied on the historical data contained in journals and in time-lapsed photography and maps although these largely are restricted to changes during the last century. There is something of an information gap between the domain of this type of record and that of reliable C^{14} dating.

Engineering time, at the high-frequency end of the time-scale continuum, is the domain of the basic science of the river engineer: fluid mechanics. Engineers traditionally have developed solutions to river problems on the assumption that the governing conditions at the geomorphic time-scale are constant. This is the basis of the grade, regime, or steady-state approach to river behaviour. Given various assumptions, fluid mechanics models provide precise quantitative information on the fluid flow-field and form of an open channel (for example, see the 1957 Rozovskii analysis of bend flow). Such models are usually developed for the time period necessary for the flow to pass through the channel reach in question. It is assumed that equilibrium channel morphology is quickly established and it follows from the steady-state assumption that long-term channel form is obtained by simple extension of a stationary series with respect to time. This assumption might be valid for short periods of time (say a couple of years or even a few decades) but it becomes less defensible as the time period in question is lengthened. Unfortunately, river engineers are commonly required to design structures with a life in the order of 100 years or more and in such cases the assumption is tenuous indeed.

Because problems of river behaviour have been perceived differently by engineers on the one hand, and by earth scientists on the other, until recently there has been little interaction among these groups of scientists. But in the last couple of decades this situation has changed markedly for at least two reasons. First, earth scientists have looked to fluid mechanics and engineering research to provide more reasoned process-oriented explanation and integration of their largely qualitative observations. This is particularly evident in sedimentology and in fluvial geomorphology. Second, engineers have looked to the earth sciences to specify the non-stationarities in their steady-state models that all too often prevent accurate forecasts of long-term river behaviour. Engineers' concerns with long-term river

adjustment to a considerable degree have been legislated by governments responding to the new environmental awareness of the public. Engineers have been required to give answers to a broad range of environmental questions, many of which are well beyond their traditional area of expertise. These social forces no doubt have quickened the pace of scientific collaboration.

Although this collaboration of river engineers and earth scientists clearly has been beneficial to both groups and to science in general, it has at the same time been a disappointing one in some respects. Engineers have gained only general insights into design problems as they relate to processes operating at geomorphic time scales and earth scientists have tended to become enbroiled in the problems of the engineering time scale without really finding these process studies very useful for explaining significant channel changes at the geomorphic time scale.

The fact is that, after a century of relatively intense effort to understand the ways of rivers, we remain unable to predict an accurate channel response to a simple change in flow regime at any time scale.

Some of the underlying problems are brought into sharper focus if we briefly review what we actually know about rivers in equilibrium at engineering time scales. These are the facts and ideas that collectively might be termed the theory of equilibrium channel morphology.

The Theory of Equilibrium Channel Morphology

Introduction

In its simplest and most general form, it states that, if left undisturbed, rivers will establish some stable combination of morphological elements for a given discharge of water and sediment. Any disturbance to this equilibrium channel morphology will set in motion processes that will return the channel to its stable form and pattern. Any change in the character of fluid discharge will set in motion changes to establish a new state of channel equilibrium in a manner dependent on the degrees of freedom of channel adjustment and on the nature of the adjustment processes.

A river conveying a given discharge possesses the potential of at least eight significant degrees of freedom for change (A.S.C.E. Task Force report, 1971c): width, depth, sediment calibre, sediment discharge, velocity, slope, boundary roughness and planform. It follows that it is necessary to specify as many processes linking these variables to fully characterise this fluvial system. The search to identify these process equations has constituted a major effort by river scientists during this century.

Three particular processes are thought to provide obvious candidate equations: flow continuity, flow resistance, and sediment transport. The exact nature of the corresponding process equations is the subject of much debate. The general character of the remaining five processes is even less well understood.

Perhaps one of the best known attempts at closure of a somewhat simpler set is that of Langbein (1964) and Langbein and Leopold (1964). They postulated two further "processes" (actually probable conditions) to the three listed above: a tendency for a river to expend equal power per unit bed area and another tendency towards equal power expenditure per unit channel length (minimum work rate). Although these two conditions allow closure and yield a supposed average hydraulic geometry, they are nevertheless, probability statements rather than process equations and a unique solution simply is not possible. Indeed, the two conditions are

mutually exclusive.

Others have attempted similar solutions (for example, see Yang, 1976, Song and Yang, 1980, Yang et al, 1981, for a minimum unit-power approach and Davies and Sutherland, 1980, and Kirkby, 1977, who respectively argue for maximisation of flow resistance and for efficiency of sediment transport) while others have concluded that the fluvial system is intrinsically indeterminate (see Maddock, 1970).

Another view, perhaps most recently stated by Hey (1978), is that the lack of any physical justification for the minimisation procedure, coupled with its inability to provide unique solutions, simply suggests that additional process equations need to be defined to obtain a determinant solution. Hey (1978) suggests that the remaining equations should specify processes of bed deformation, bank erosion, and of meandering. Few would argue about the importance of these processes but of course it is one thing to state the nature of the problem and altogether another to solve it.

It is useful to our developing a sense of the state of the fluvial art to consider some of the better defined process equations.

The Process Equations

Flow continuity is a basic physical relationship which can be expressed as

$$Q = w\bar{v} \quad (1)$$

where Q is discharge, \bar{v} is mean flow velocity, and w and \bar{d} are respectively the width and mean depth of the channel. This fundamental definitional equality, however, is deceptively simple. Although the continuity equation finds relatively direct application in laboratory experiments with controlled flows, highly unsteady natural flows are not conveniently reduced to a single summary discharge. The width and mean depth of a river channel clearly are determined by all discharges greater than that corresponding to threshold sediment transport, integrated over a time period limited by the rate of lateral or vertical channel movement. But most river scientists recognise that not all discharges in this domain are of equal effectiveness in forming a river channel; a far more limited range is thought to be dominant.

Although the dominant discharge concept has become firmly entrenched in fluvial geomorphology and hydraulic engineering, it remains a controversial idea (see Riley, 1972; Williams, 1978). Many researchers agree with Dury (1960, 1961, 1965) that, not only is the concept of a single dominant discharge valid, but that it can be equated to bankfull discharge (see Leopold, Wolman and Miller 1964, and the more recent discussion of Ackers and Charlton, 1970). Furthermore, Dury and others embraced Nixon's (1959) implied notion that bankfull discharge corresponds with the most probable flood on the annual series (Dury, 1961; Dury, Hails, and Robbie, 1963). Many researchers would argue, however, for several dominant discharges (Woodyer, 1968) or for a limited range of dominant discharge (Carlston, 1965). More recent work has developed the idea that the frequency of bankfull discharge and dominant discharge may differ and that both are dependent on the type of river regime (see Benson and Thomas, 1966; Harvey, 1969; Pickup, 1975; Pickup and Warner, 1976. and Pickup and Rieger, 1979).

Difficulties of applying the dominant discharge concept will be compounded if it is shown that different components of channel morphology (say mean depth and meander wavelength) scale with different parts of the discharge distribution (see Neller, 1980; Rood, 1980). One thing clearly is certain: the basic problem of

specifying dominant discharge in rivers remains unresolved.

Flow resistance, the second of the three process equations, constitutes a still more elusive relationship although a commonly adopted empirical formula is

$$1 / \sqrt{ff} = v/V_* = C_1 (\bar{d}/k_s)^{C_2} \quad (2)$$

in which $V_* = \sqrt{g\bar{d}s}$, the shear velocity, and ff is the Darcy Weisbach resistance coefficient. If the Strickler relation $n = 0.034D^{1/6}$ for the resistance affect of grain size, D , applies, then $k_s = D$, $C_1 = 8.4$, and $C_2 = 1/6$ in equation (2), yielding the popular Manning equation.

Although equation (2) has enjoyed successful application in channels with regular fixed boundaries, it becomes increasingly difficult to apply when allowances are made for the varied character of natural alluvial channels (see A.S.C.E. Task Force report, 1963). For example, although we now better understand the nature of stability criteria for bedforms than we did two decades ago (for example, see Simons and Richardson, 1963, 1966; Simons, Richardson and Nordin, 1965; Rathburn, Guy and Richardson, 1969; Southard, 1971; Middleton and Southard, 1977; Yalin, 1977; Dalrymple, Knight and Lambiase, 1978; Allen and Leeder, 1980) we remain unable to predict the contribution of bedform roughness to k_s in equation (2) (see Shen, 1973, and Simons and Senturk, 1977, for a review of many such attempts including the pioneering work of Einstein and Barbarossa, 1952). It appears that our ability to do this is some way off, depending as it likely does on a far more comprehensive understanding of macroturbulence structure (particularly at high discharges) than we presently can claim (see Matthes, 1947; Demente'ev, 1963; Popov, 1964; Korchokha, 1968; Grass, 1971; Kim et al., 1971; Laufer and Narayanan, 1971; Grinval'd, 1972; Jackson, 1975, 1976; Nowell, 1975; Nowell and Church, 1979; McLean and Smith, 1979; Rood, 1980).

Resistance to flow, however, clearly is also dependent on many factors other than grain and form roughness (see A.S.C.E. Task Force report, 1963; Simons and Senturk, 1977); not the least of these is the influence of channel bends. But here again, although the the general role of planform in determining flow resistance has been recognised for some time (see Bagnold, 1960; Leopold et al., 1960) specification of the exceedingly complex nature of bend flow remains beyond our present capability (see Callander, 1978).

It is reasonable to conclude that a general predictive flow resistance equation is yet to be developed!

Sediment transport, the third process equation, is of course the subject of a vast literature. Although any sort of review of this literature obviously is beyond the scope of these introductory comments, it is useful to consider the nature of a few general problems.

The role of sediment transport in determining channel morphology is set in context by equation (3):

$$\frac{\partial z}{\partial t} = \frac{\partial G_s}{\partial x} \quad (3)$$

Here the rate of change in bed elevation on the left of equation (3) is dependent on downstream changes in the sediment transport rate. The transient solution of

equation (3) is the solution of the degradation/aggradation problem and has been attempted by several researchers (e.g. see Tinney, 1955; Culling, 1960; Scheidegger, 1965, 1970; Devdariani, 1967; Soni, 1981). Central to all of these solutions is the specification of the sediment transport function.

Most of the widely used equations designed to predict the important bed load component of G_s in equation (3) are a variant of the early Du Boys relation in which the rate^s of bed load transport is expressed as a simple function of the excess shear stress beyond that necessary to initiate sediment movement. These Du Boys-type equations can be expressed in terms of threshold shear stress or in terms of threshold velocities or discharges but all require specification of the critical flow conditions for sediment movement. Several important equations of this type bear the names of their originators (Shields, Schoklitsch and Meyer-Peter and Muller) and are the subject of discussion in most reviews (see Raudkivi, 1967, A.S.C.E. Task Committee, 1971a, 1971b; Graf, 1971; Bogardi, 1974; Simons and Senturk, 1977).

The only significant major departure from the mean tractive force concept used by Du Boys is the work of Einstein (1942, 1950). His partly stochastic approach, developed from concepts of fluid turbulence, assigned a probability for movement to each bed particle, thus avoiding the use of a threshold entrainment condition. Einstein's "bed load function" and functionally similar equations (see discussion by Chien, 1956) proposed by others such as Kalinske (1947) and Graf and Acaroglu (1968) are discussed in the reviews cited above.

It should be noted that, although most of these sediment transport equations have a general form based on theoretical considerations, all include important empirically determined coefficients largely based on data from laboratory flumes. This fact immediately raises the question of the appropriateness of these laboratory-determined constants in a field situation. In most circumstances such questions are simply resolved by field observation but unfortunately this case is further complicated by problems of measurement. Except on very small rivers or at unusually elaborate control structures, it is extremely difficult to obtain an accurate measurement of bed-load transport. Our present instrument technology is simply inadequate in most cases. Nevertheless, it seems likely that the several orders of magnitude range in bed-load-transport estimates provided by the general-use equations (see Vanoni, 1975) implies that many of the assumed constants are in fact unspecified variables.

One of the primary culprits is the universal assumption that the immersed weight of a bed particle alone determines the force necessary to initiate its movement (or determines its probability of movement in the stochastic models). It may be a reasonable assumption in a laboratory flume with spherical bed particles but it is a highly dubious one in many natural channels. For example, it ignores armouring effects and sheltering effects of imbrication in gravel-bed streams (see Harrison, 1950; Livesay, 1965; Kellerhals, 1967; Gessler, 1970, 1971; Church and Gilbert, 1975; Day, 1976; Little and Mayer, 1976; Bray and Church, 1980) and ignores the effects of organic mats that can form a protective coating during low flows in sand-bed channels.

The bed armour problem is really part of the broader fundamental question of whether equilibrium sediment transport conditions are commonly attained in natural rivers. All sediment transport equations attempt to predict the maximum sediment transport rates given certain fluid dynamic conditions. That is, they assume that sediment supply is not a factor limiting sediment transport. There is more than a little evidence to suggest, however, that this simply is not true. The hysteresis

effect in sediment concentration associated with the initial flushing of sediment during passage of a flood wave is a widely recognised phenomenon. Variations in the supply of freeze-thaw debris are known to produce seasonal variations in sediment transport rates in some rivers (see Nanson, 1974). An example of the analogous effect at a geomorphic time scale is provided by Church and Ryder (1972). They describe the Canadian landscape as one of declining intensity of geomorphic activity since immediate deglaciation. Although sediment supply from the freshly glaciated land surface to rivers is unlikely to have been a factor limiting sediment transport rates in immediate deglacial times, much of this glacial material has now been flushed from the geomorphic system and the sediment supply rate has declined. This fact together with the tendency for Canadian gravel-bed rivers to develop armoured beds, suggests that many of them may well be transporting sediments at a rate which is supply limited.

In the case of sediment-supply limited transport rates, sediment yields from drainage basins must be known if these rates are to be estimated. It is sufficient to say that, in this realm, our understanding at best is qualitative and completely inadequate for the task at hand.

Perhaps I should leave the final remarks on sediment transport to Simon and Senturk (1977) who conclude that "The mechanics of sediment transport is so complex that it is extremely unlikely that a full understanding will ever be obtained. A universal sediment transport equation is not and may never be available" (p. 644).

The equations for closure, as indicated earlier, are not self-evident but certainly must include (in addition to the three so far considered) a governing equation for lateral movement of channels. Such movements include the complex set of sinuous motions that we collectively term meandering.

Meandering is a significant process at both engineering and geomorphic time scales. Lateral migration is sufficiently rapid on many large rivers that it can result in the channel shifting sideways by a complete channel width in less than 20 years. It therefore is a process that can change mean water-surface slopes rapidly (within the limits set by valley slopes) and it can cause local catastrophic changes through the formation of cut-offs. Meander development also introduces considerable flow distortion at bends which must be incorporated into any comprehensive flow resistance equation. It is also a significant factor governing local sediment supply.

Unfortunately, meandering also is a process about which we have a great deal to learn. Although some general empirical relationships among discharge and various planform elements of meandering channels have been known for some time (see Leopold-Wolman, and Miller, 1964; some of these will be discussed later in this introduction), the underlying causes and controls of these regularities are yet to be identified.

Empirical studies of planform geometry apart, research into meandering has been conducted largely independently on three closely related topics: the cause of meandering, the nature of bend flow, and the nature of lateral migration.

The cause of meandering, a fundamental and long standing question in fluvial studies, remains unanswered to this day. Earlier work, in which helical flow is featured as an important process, is briefly reviewed by Leopold, Wolman and Miller, (1964) and more recent developments are discussed by Callander (1978). Currently in favour is the case for dynamic instability of the alluvial channel-bed. Stability analyses have been used to argue (see Callander, 1969; Sukegawa, 1970; Engelund and Skovgaard, 1973; Engelund, 1974; Hayashi, 1974; Parker, 1976; Ponce and Mahmood, 1976) that, because the mobile bed of a straight channel is unstable, any small

perturbation will be amplified to produce the pattern of alternating pools and riffles that Friedkin (1945) and many others subsequently have noted are the precursors to meandering. Although these rather complex mathematical models have certain basic elements in common (periodic functions for the initial perturbation in which the wavelength of maximum amplification is assumed to be the corresponding meander wavelength) they employ a variety of other assumptions to close the set of flow equations. In spite of the variety of specific solutions, all succeed in predicting wavelengths similar to those found in natural channels! It is not clear whether this is a weakness or a strength of this approach! Certainly the general nature of the approach is appealing; the periodic function does not specify the process of perturbation and therefore can easily accommodate meandering tendencies in a variety of media. Parker (1976) argues that, although perturbations in sediment transport are necessary to form meanders in rivers, variations in Coriolis acceleration serve the same purpose in oceanic currents, heat differences in meltwater streams, and surface tensions in water threads on glass surfaces. This generality at the same time leaves stability analyses open to the criticism of being "black box" solutions to complex problems.

Other explanations of meandering proposed in the last couple of decades are less general than the stability analyses but are just as lacking in specific process information. For example, Shen and Komura (1968) and Quick (1974) appeal to a periodic reversal in vorticity while Yalin (1971, 1977) argues that meander wavelength scales with the largest macroturbulent eddy that will fit in the channel. The debate continues.

Once a channel bend begins to develop, for whatever reason, the flow is characterised by the interaction of two sets of forces: those demanding the conservation of angular momentum (free vortex flow) and those promoting the lateral transfer of momentum (helical flow). The resultant pattern of bend flow is directly related to erosion and deposition in the channel and is the primary control on the rate and pattern of lateral migration.

Theoretical studies of bend flow, in which solutions have been sought for the continuity and momentum equations for turbulent flow, have led to successful prediction of its general character. Early research is reviewed by Rozovskii (1957) and later contributions are discussed by Yen (1965) and Callander (1978). Recent advances by Yen (1972), Engelund (1974), and Kikkawa, Ikedo and Kitagawa (1976) have improved the theory to the point that it can provide realistic descriptions of the three-dimensional flow field (including the correct sense of helical flow and secondary flow strength of the right order of magnitude), of the bed shear-stress distribution through the bend, and of the general bed configuration. It clearly has been one of the more successful areas of theoretical analysis although the theory certainly is not without rather significant limitations. For example, it only applies to bends with a single helical cell and perhaps more importantly, it does not apply to bend flow in which there is flow separation. Although the normal presence of more than one helical cell remains debatable, the occurrence of separation zones is widely recognised to be a feature common to most experimental and field observations of bend flow (see Callander, 1978).

Although Yen (1975) and Allen (1977) have offered general analytical comment on bend erosion, a theoretical model of lateral migration is not available.

The pace of experimental research on bend flow has increased markedly since the classic works of Mockmore (1943) and Rozovskii (1957) to peak in the mid 1970's. Some of the more notable contributions include those of Einstein and Harder (1954), Ippen and Drinker (1962), Yen (1967, 1970), Fox and Ball (1968), Yen and Yen (1971),

Francis and Asfari (1971), Martvall and Nilsson (1972), Hooke (1974, 1975), Engelund (1974, 1976), Ikeda (1974), Monsonyi and Goetz (1974), Monsoyi et al. (1975), Meckel and Veng Heang Chhun (1975), Siebert and Gotz (1975) and Varshney and Garde (1975). Although these experimental studies provide test data for the theoretical models and have added to our understanding of bend-flow processes, they have not yielded a quantitative empirical framework for erosion prediction. Indeed, in this context many of them may have provided misleading results (Hickin, 1977). Few experimental studies are of live-bed conditions and most adopt constant-width channels with rectangular or trapezoidal sections, a simple geometry which likely represents a significant departure from reality.

There remains a gap between experimental studies of this fixed-bend type and those that have attempted to characterise the initiation and growth of meander bends (see Friedkin, 1945; Wolman and Brush, 1961; Hickin, 1969, 1972; Schumm and Khan, 1972). The general pattern of alternate bar formation flow deflection, and the alternating bank erosion that leads to development of a sinuous channel have been known for some time. But no experiment has related the development of this sinuous deformation of the channel alignment to the changes in the velocity and shear distributions in the flow. Furthermore, most experiments have been conducted in non-cohesive sediments, producing pseudomeanders (see Wolman and Brush, 1961) rather than analogues to natural meanders with pools and riffles. The more realistic results of Schumm and Khan's (1977) experiments using a kaolinite/sand sediment have highlighted this distinction in meander type.

There are still fewer studies of flow in real river bends. The remarkable exposition of fluvial processes on the River Klaralven by Sundborg (1956) might have established an early trend but it has taken until the 1970's for his lead to be followed. Jackson (1975a, b, 1976) measured primary and secondary flow velocities around several bends of the Lower Wabash River in order to better understand the pattern of point bar sedimentation there. Bridge and Jarvis (1976) and Bridge (1977) have described primary and secondary flow velocities around a bend of the River South Esk in Scotland. They extended the theoretical concepts of Rozovskii (1957), Allen (1970a, b, 1971) and Engelund (1974a, b, 1975) to develop a three-dimensional model of bend sedimentation. Hickin (1978) obtained measurements of primary and secondary flow velocities through a continuous series of bends on the gravel-bed Squamish River in British Columbia. It is the only field study specifically designed to yield information on the hydraulic control of lateral migration rates.

These studies confirm the main theoretical and experimentally based generalisations about bend flow but they also point to important factors ignored in this earlier work. Principal among these is the interaction of flow in adjacent bends and the inevitable occurrence of separation zones (also see Carey, 1969; Woodyer, 1975; Taylor et al., 1971; Leeder and Bridges, 1975; Hickin, 1979; Nanson and Page, 1981). Unfortunately, these are factors of great significance to the lateral migration process and will have to be understood before an appropriate model can be developed.

The reality is that a universal channel migration model probably never will be available. It is after all just another version of the sediment transport problem further complicated by the exceedingly more complex flow and entrainment conditions involved. These circumstances make actual measurements of the process vital to our ability to forecast channel changes.

There have been very few direct field surveys of channel migration; Lewin (1977) provides a brief review of some recent efforts. The successive surveys of Watts Branch, Maryland, first reported by Wolman (1959) and later by Leopold (1973) provide a well known example of this type of direct monitoring. Other direct

observations of channel shifting have been reported by Coleman (1969), Daniel (1971), Hughes (1977), Lewin and Brindle (1977), and Hooke (1977, 1980).

Direct monitoring of channel changes at monumented transects is a recent development and one which, because of the considerable maintenance effort required and the uncertainty of obtaining useful results, is rarely sustained for more than a few years. It follows that the method has been most useful for describing channel migration processes at relatively short time-scales of a few days to just a few years at most. Although this type of direct measurement provides valuable information on the process of lateral migration, the sampling time is far too short to provide a stable mean migration rate for time scales of the order of decades or longer.

The great majority of lateral migration measurements, however, have been obtained indirectly from serial cartography and aerial photography of channels. This type of historical record generally extends the period of observation to a few decades in the case of aerial photographs and up to a century or so in the case of cartography. Clearly, its usefulness will vary considerably from place to place. In areas with a long cultural history the cartographic record is likely to be useful (at least five surveys over the last 130 years are generally available in Britain (see Lewin and Hughes, 1976) but it is minimally useful in much of the New World. The detailed record of channel pattern for parts of the Mississippi-Missouri system in the United States (Mississippi River Commission, 1939, 1945; Carey, 1969; Ruhe, 1975) provides a notable exception to this generalisation.

Systematic photogrammetric surveys generally are restricted to the post-war period although specific projects may have been undertaken in the first half of this century in some areas. For example, a typical maximum time interval for the photographic recording of channel migration in most of Canada and in much of the United States is about 30 years. Except in special cases, most areas are not likely to be aeriually photographed any more frequently than once every five or ten years; this practical limit of coverage defines the short-term limit of resolution of channel migration processes.

This type of comparative information from maps and aerial photographs is rapidly and cheaply obtained and is routinely collected on a project by project basis by a variety of agencies in many countries.

Channel migration rates for time periods from about one to five hundred years have been obtained from the analysis of flood-plain vegetation successions (see Eardley, 1938; Everitt, 1968; Hickin and Nanson, 1975). Nanson (1977) and Nanson and Beach (1977) also report associated rates of vertical flood-plain accretion based on an analysis of wood-cell structure in cottonwoods. The lower limit of resolution of flood plain dating by dendrochronology is set by variability in tree-colonisation times (probably uniformly small for most species) and subsequent survival rates; the length of the record obviously is dependent on the life of the tree.

Few generalisations can be made from these recent field studies that have not already been stated in the classic work of Leopold, Wolman and Miller (1964). Perhaps one of the more significant developments has been the willingness to recognise the wide variety of bend evolution other than simple expansion to a sine-generated form and its subsequent downvalley translation. A number of case studies have stressed the complexity of combined expansion, rotation, and translation in bend development (see Schmudde, 1963; Makkavyeyer et al., 1969; Daniel, 1971; Handy, 1972; Chitale, 1973; Hooke, 1977). A very common meander form is the complex compound meander loops described by Brice (1974a, b) and Hickin (1974). Hickin and

Nanson (1975) confirmed that migration rates at the apex of bends on the tortuous Beatton River in British Columbia are dependent on channel curvature which plays an important but not precisely identified role in governing the development of complex meander lobes.

Another noteworthy development in recent studies of river planform is the increased recognition of the importance of cut-offs in meander kinematics (see Vogt, 1963; Speight, 1965b, Kondrat'yev, 1968; Neill, 1970; Kulemina, 1973; Schäfer, 1973; Mosley, 1975a, b). Until the last few decades our preoccupation with "equilibrium" forms has led to the dismissal of cut-offs as transient disturbances to the fluvial system. In fact, cut-offs form frequently enough on many rivers that certain reaches may remain in this transient state permanently!

One of the preliminary results of a recent investigation in Western Canada (Nanson and Hickin, 1981) is that mean migration rates over a 25-year period (from aerial photographs) may bear little relation to that for a century or longer (from the vegetation record). Initial findings suggest that the photographic record is too short to fully integrate a low-frequency (20-30 year periodicity?) pulsing in the rate of lateral migration. This pulsing may relate to the fact that the rate of bank erosion is governed by higher magnitude and lower frequency floods than those that govern the rate of lateral accretion on point bars. This difference in rates of lateral movement causes the channel width in the bend to fluctuate; when it is narrow, short-term migration rates are high and when it is wide they are relatively low. If this generally is so, almost all measurements of migration behaviour based on aerial photographs are overwhelmingly biased estimates of longer-term behaviour. This hardly can be a comforting thought for the river engineer!

Some Conclusions

It is clear from this brief review exercise that in each process case considered, our knowledge is inadequate to define a set of general governing equations that together would constitute a theory of equilibrium channel morphology at an engineering time scale. For certain purposes, however, the continuity equation for rivers may be relatively well defined and the resistance equation may be an adequate approximation. There also remains some prospect that further significant advances will be made in our understanding of the lateral shifting of rivers. But the existing sediment transport equations are simply inadequate in almost all cases. Not only do they apparently not accurately describe the mechanics of sediment movement but it appears likely that the flow-limited equilibrium sediment transport rate they are designed to predict is both rare and insignificant in shaping real channels. It seems that there may be a far better prospect of predicting sediment transport rates from channel morphology rather than vice versa! Perhaps it is time to admit that accurately estimating sediment transport rates in rivers using a fluid mechanics approach has been demonstrated to be a practically insoluble problem. Some would argue that we have been preoccupied with this approach to the sediment transport problem for quite long enough; our research resources could be more usefully devoted to other causes. Certainly we must conclude that process-based theory for predicting channel changes simply does not exist.

This general conclusion, of course, is hardly a novel one. It was this same conclusion made a century ago that persuaded river engineers to develop the empirical sets of norms that we have come to know as regime theory and hydraulic geometry.

Regime Theory and Hydraulic Geometry

Empirical studies have a long tradition both in the earth sciences and in

engineering research. The engineering work on the stable alluvial irrigation canals on India marks an appropriate beginning to such river studies in the modern era. Extending the early observations and ideas of R.G. Kennedy (1895), Gerald Lacey (1929, 1933, 1946, 1958) developed a set of empirical relationships describing the morphology of stable alluvial canals for a limited range of discharges, water-surface slopes, and of a sediment coefficient (now known as the Lacey silt-factor). Although these regime relationships adequately describe the geometry of Punjab canals, they have not found a similar degree of success in wider applications. The coefficients of Lacey's empirical equations clearly reflect fortuitous constancy in certain parameters which are apparently important variables in other fluvial contexts. For example, E.W. Lane was prompt in pointing out in 1937 that the Lacey silt factor does not appear to adequately account for variations in boundary materials and sediment loads. Subsequently, Inglis (1948) and perhaps the best known of contemporary regime-method engineers, Tom Blench (1966), have attempted to incorporate sediment load parameters into the regime equations. Nevertheless, they remain empirical equations which only will apply well to the limited set of conditions from which they were derived. A discussion of the relation between regime theory and critical tractive force theory is provided by Henderson (1961); a general review of regime equations is given by Mahmood and Shen (1971).

Regime theory is a very significant contribution to the study of channel adjustment for a variety of reasons. It continues to be a widely used design tool obviously judged by many engineers to be the best technique available for predicting stable channel size and form. Also, it rests firmly on the concept of channel equilibrium, one which we have noted is fundamental to any discussion of river channel changes. Perhaps most important is the fact that, with the engineering steady-state concept of equilibrium relaxed (quasi-equilibrium), regime theory provided the foundation and inspiration of the general geomorphic discussion of river form in the 1950's and '60's that to earth scientists became known as the hydraulic geometry of stream channels.

The hydraulic geometry concept is discussed in some detail by Leopold, Wolman and Miller (1964) reviewers who contributed much to its development. Briefly, it is the set of power relations between discharge and principally the width, mean depth, and mean flow velocity, but also the water-surface slope, boundary roughness and sediment load, of the natural channel. Of the two basic types of hydraulic geometry, at-a-station and downstream, it is the latter that has application to problems of river channel change. At-a-station hydraulic geometry in most cases simply describes the way in which changing discharge fills the effectively rigid-boundary channel that has been moulded by an earlier and larger discharge. The downstream hydraulic geometry, on the other hand, describes how the fully adjusted channel form accommodates changes in a competent bankfull (or near bankfull) discharge; it is in this "downstream" context that hydraulic geometry and regime theory are essentially one and the same view of channel equilibrium.

It is important, however, that we not lose sight of the fact that the hydraulic geometry simply describes what is there. It has little or no theoretical significance. Even the power function form ascribed to the relations is done as much for convenience as for any other reason. There is no one hydraulic geometry and it is most useful for engineering purposes when it is developed for an environmentally homogeneous region (for example, see Bray, 1972, 1973, 1975, 1979, 1981).

Similar comment can be offered on the empirical relationship, $l = c q_{df}^{1/2}$, widely acknowledged to exist between average meander wavelength (l) and bankfull discharge (q_{df}). The fact that meander geometry appears to scale with channel width and with the square root of discharge was recognised by a number of researchers including

Inglis (1949), Friedkin (1945), and Leopold and Wolman (1960), but it was George Dury who provided the most comprehensive treatment of its significance as an indicator of environmental change. In a series of papers (1953, 1954, 1955, 1958, 1960, 1964a, 1964b, 1965 and 1976) he exploited the meander wavelength/discharge relationship to estimate the discharge reductions and associated Quaternary climatic changes implied by regional underfitness of rivers. But even at this level of generalisation many researchers have not been persuaded that the meander wavelength/discharge relation is sufficiently well-defined to warrant these deductions about the former regional climatic and hydrologic environment.

Much of the criticism has focussed on three issues. The first concerns the adoption of a single summary discharge variable to characterise the formative flow and is a concept central to both the downstream hydraulic geometry and to Dury's meander planform analysis. The problem is detailed here in the earlier discussion of the continuity equation.

The second and similar issue concerns the appropriateness of the simple average meander wavelength as the index of meander scale. It has been argued that the variation in meander wavelength in a given reach of river does not represent normally distributed excursions from some mean wavelength but rather reflects the interaction of several fundamental wavelengths related to discharge variability. It was this view of meander planform that led Speight (1965a, b, 1967) to apply spectral analysis to meander arrays in order to express the contribution of the assumed component harmonics to the resultant waveform. Other researchers have pursued this line of enquiry (Chang, 1969; Thakur and Scheidegger, 1970; Chang and Toebes, 1970; Church, 1970, 1972; Yalin, 1971; Ferguson, 1975, 1976; Hickin, 1977) although the collective results so far have been less than conclusive.

The third issue concerns the general question of the influence of boundary materials on channel form and pattern. For example, Dury (1977) recounts his case for there being no difference between bedrock and alluvial meanders and yet this seems contrary to the principle that boundary material influences the width of a channel which in turn is a scaling factor for meander wavelength (Shahjahan, 1970).

Although boundary material variation usually is statistically unimportant in these bivariate relationships in which discharge ranges over several orders of magnitude, it becomes increasingly more important as the discharge range contracts. In the limit, knowing the discharge allows a mean morphology to be specified but the wide scatter of possibilities about this mean condition impose a no better than order of magnitude level of resolution.

It was on the assumption that much of this scatter could be attributed to variation in boundary materials that Schumm based his empirical studies of channel morphology, discharge, and bed and bank materials (Schumm, 1960, 1963, 1965, 1968, 1969, 1972). In a sense Schumm's work marks a return to the approach advocated by the regime theorists who earlier attempted to incorporate sediment type into their models. His work is the most widely adopted set of empirical relationships used for palaeohydrologic reconstruction although the approach is not a new one (see Lane, 1955; Blench, 1972).

Schumm collected data on flow and channel characteristics from rivers on the Great Plains of the United States and on the Riverine Plains of Eastern Australia. Statistical analysis of these data yielded the following regression equations:

$$w = 37Q^{0.38}/M^{0.39} \quad (4)$$

$$\bar{d} = 0.6M^{0.34}Q^{0.29} \quad (5)$$

$$F = w/\bar{d} = 255/M^{1.08} \quad (6)$$

$$M = 55/Q_{bs} \quad (7)$$

$$S = 60/M^{0.38}Q^{0.32} \quad (8)$$

$$\lambda = Q_{ma}^{0.48}/M^{0.74} = 1890 Q_m^{0.34}/M^{0.74} \quad (9)$$

$$P = 0.94M^{0.25} \quad (10)$$

in which Q = discharge (ft^3/sec), Q_m = mean annual discharge (ft^3/sec), Q_{ma} = mean annual flood (ft^3/sec), Q_{bs} = bed load as a percentage of total load, w = channel width, \bar{d} = channel mean depth (ft), S = water-surface slope, λ = meander wavelength (ft), P = sinuosity and M = the percentage of silt-clay in the perimeter of the channel and an index of sediment load.

Equations (4) and (5) include a sediment parameter which statistically explains much of the residual variation in channel width and depth among the rivers in question left unexplained by variations in discharge. Similar improvement in the definition of the "meander law" is achieved by taking account of sediment type in equation (9).

For the rivers from which these data were obtained, this closed set of equations provides an empirical model for predicting channel changes. If the direction of these relationships is universal, and if bedload transport (Q_s) is inversely related to the sediment parameter, M , equations (4) to (10) imply that^S

$$Q \propto \frac{w, d, \lambda}{S} \quad (11)$$

$$\text{and } Q_s \propto \frac{w, \lambda, S}{d, P} \quad (12)$$

These general relationships lead to the following set of channel-change associations (see Schumm, 1977):

$$Q^+ \propto w^+, d^+, \lambda^+, S^- \quad (13)$$

$$Q^- \propto w^-, d^-, \lambda^-, S^+ \quad (14)$$

$$Q_s^+ \propto w^+, d^-, \lambda^+, S^+, P^- \quad (15)$$

$$Q_s^- \propto w^-, d^+, \lambda^-, S^-, P^+ \quad (16)$$

$$Q^+ Q_s^+ \propto w^+, d^+, \lambda^+, S^+, P^-, F^+ \quad (17)$$

$$Q^- Q_s^- \propto w^-, d^-, \lambda^-, S^-, P^+, F^- \quad (18)$$

$$Q^+ Q_s^- \propto w^+, d^+, \lambda^+, S^-, P^+, F^- \quad (19)$$

$$Q^- Q_s^+ \propto w^-, d^-, \lambda^-, S^+, P^-, F^+ \quad (20)$$

Equations (13) to (20) describe a number of possible morphological responses to a variety of changes in the discharge of water and sediment, some more likely than others. Santos-Cayade and Simons (1973) and Schumm (1977) provide a discussion of circumstances in which these changes might occur. Clearly, even for this

qualitative scheme, many of the changes are indeterminate, particularly for the more realistic cases, because the magnitude of opposed responses are unspecified. In such circumstances it is very tempting to solve this problem by making the completely unwarranted assumption that equations (4) to (10) are universal. Herein lies the fundamental limitation of this empirical approach: the lack of field data.

Given the difficulties associated with the process-oriented approach discussed earlier, this empirical approach deserves far more attention than it has received. Oddly enough, it is almost the only study of its kind in spite of the fact that it was completed over two decades ago. There seems to be an inexplicable aversion on the part of earth scientists and engineers to repeat the field experiments of others.

The danger of generalising from particular empirical norms is well illustrated by considering one of Schumm's (1960) central relationships: that between the ratio width/depth of a channel and the weighted percentage silt-clay content of the bed and bank materials (equation 6). This relationship is based on 69 channel sections on about 30 rivers in the American Midwest. Today we still have little idea whether or not this relationship applies to other Midwestern rivers or to any in regions beyond. But of course we are very willing to assume that it does. We also seem to have some sort of intellectual defence mechanism that prevents us from recognising the order-of-magnitude scatter in this particular relationship. For example, it hardly can be comforting for an engineer to know that boundary materials with 10 percent silt-clay content are associated with channels having width/depth ratios that commonly vary between 10 and 60! Perhaps it is even more disturbing that at least one independent field check of this relationship (65 river cross-sections from the Namoi-Gwydir distributary system of Eastern Australia) has shown that equation (6) does not apply there at all (Riley, 1975).

These observations do not invalidate the empirical approach but they do point to the need for a more comprehensive data set than presently is available. The norms would be much more useful as a planning and design tool if data were collected and analysed for geologically and hydrologically homogeneous areas. Such a geographic exercise would strengthen the order in particular hydraulic geometries while recognising the variety in general. The regional regime relations for Albertan gravel-bed rivers developed by Bray (1972, 1973, 1975, 1979, 1981; also see Kellerhals, Neill and Bray, 1972) is a good example of this type of exercise.

Probably the single most important limitation of the approach of hydraulic geometry, one that it shares with the theory of equilibrium channel morphology as previously discussed, is that it only describes static equilibria. The process of channel adjustment from one equilibrium state to another generally is assumed to be instantaneous; transient states are not accommodated although they appear in the hydraulic geometry as scatter in the data (Knighton, 1975, 1977). In other words, rates of change in channel morphology remain largely unspecified. Even if a new equilibrium state can be specified, it is not possible to predict how long it will take to reach it.

Related to this problem of specifying rates of channel change is that of accounting for the effects of large floods. Extreme events, part of the normal process system, may cause significant departures from equilibrium channel morphology that may persist for long periods of time (relaxation time). The Schumm and Lichty (1963) study of the Cimarron River in southwestern Kansas provides a well-known example of this behaviour. They reasoned that a large flood in 1914, which severely eroded the floodplain, caused a considerable influx of coarse sediment, resulting in the widening and steepening of the channel between 1914 and 1942. Since that time the channel has slowly worked towards reestablishing its pre-flood morphology

(w/d halved between 1939 and 1954) Burkham (1972) describes a similar history for the Gila River in Arizona. Similarly, many coastal rivers in eastern New South Wales have channel morphologies which appear to reflect the effects of severe regional flooding there in 1949 (Hickin and Page, 1971). Schumm (1977) cites several other examples of flood-related channel adjustments.

There may be many more examples of flood-dominated channel morphology that have gone unrecognised. Our recent thinking in this respect has been greatly influenced by the notion that a large proportion of the work done by rivers is effected during events of modest magnitude and relatively high frequency (Wolman and Miller, 1960; Gupta and Fox, 1974) although the question recently has been reconsidered by Dury (1977, 1980). It seems likely that many rivers confirm this conventional view at geomorphic time-scales but, like the Cimarron River, are characterised by long periods of flood-related transient behaviour at engineering time scales (also see Mosley, 1975; and Stevens et al., 1975; and Graf, 1981).

Another important limitation of both the theoretical and empirical approaches to channel changes is the existence of non-linearities in channel adjustment. Some of these non-linearities in relatively low flows (for example, the onset of sediment transport) and in intermediate flows (for example, bed-form changes) have been described (see Cullbertson and Dawdy, 1964; Richards, 1973) but there also may be as important ones at the often little measured but very significant high flows (for example, those related to the possible onset of general vortex shedding from the boundary). There is a need to collect detailed data sets that will allow these types of non-linearities in the hydraulic geometry to be distinguished from the background noise.

A far more profound set of non-linearities, however, is embodied in the concept of complex response in channels (Schumm, 1977).

Complex Channel Response

These channel responses include all those which are more complex than a simple direct cause and effect relationship. They involve the ideas of system thresholds (Schumm, 1973) and of episodic development of the fluvial system (Schumm, 1976). These channel responses are governed by an exceedingly complex and poorly understood set of dynamic relationships which largely have been discussed at a conceptual rather than operational level (Hey, 1979).

An interesting and probably very significant example of complex channel response is the experimental work of Parker (1976) conducted in the rainfall erosion facility at Colorado State University (see also the much earlier and simpler experiments reported by Lewis, 1944). Parker measured sediment yields from an experimental basin immediately following simulated channel rejuvenation by a lowering of base level. As the erosional knick point moved headward through the drainage network, sediment yield at first increased and then declined (as valley storage capacity increased) and subsequently increased again as major tributaries became rejuvenated, only to decline again, and so on. Sediment yields fluctuated over time because of variation in sediment supply relative to storage capacity. A consequence of this fluctuation is that downstream reaches of the experimental channel were at times aggrading (high sediment supply) and at others incising (low sediment supply), all in response to a single lowering of base level while all other external factors (climate/discharge) remained constant. Schumm and Parker (1973) believe that such a complex-response mechanism goes far in explaining the lack of correlation among terraces between and even within valleys in the American southwest (Kottlowski et al., 1965). The conventional assumption of externally-governed and therefore regionally

in-phase terracing no longer appears to be necessary. Also, there is no reason to suppose that such a process is not significant across all timescales of channel change. Again we see evidence that significant channel change is governed by sediment supply, strengthening an earlier argument made here in the context of sediment transport theory. It would seem judicious, however, that these ideas receive far more experimental confirmation than they currently enjoy before being embraced as matters of principle.

Another of the more important ideas of complex channel response is the exceedance of critical limits or thresholds that can lead to a complete transformation of river morphology (river metamorphosis: Schumm, 1969, 1971). An oft-cited example of this process is that implicit in the various channel pattern stability domains defined in the slope-discharge plane by Lane (1957), Leopold and Wolman (1957), and Ackers and Charlton (1971). These observations suggest that, at a constant discharge, if channel slope is slowly increased, straight channels will be the stable pattern at first but will give way to meandering, and finally braided channels represent the stable pattern at high slopes. At the limits of the domains, very small changes in slope can alter the stable pattern from, for example, meanders to braids. Schumm and Khan (1972) experimentally confirmed this sequence of transitions although there is no agreement on the precise location of the domains on the slope-discharge plane. It is likely that this simple bivariate plot shows association rather than process and that a more consistent differentiation of pattern would be achieved in a scheme including boundary materials and sediment transport rates (see Henderson, 1961).

It is a curious fact that field confirmations of the original Leopold and Wolman (1957) slope-discharge relationship for braids and meanders are not available in the literature. In this case, as in too many others, early ideas remain speculative because of the failure of earth scientists to follow through with the collection of appropriate field evidence. This has not prevented, however, some of these ideas becoming the foundation of further deductions.

Schumm (1977) has suggested that the many documented cases of channel straightening and widening in the United States are responses to increased peak flows and sediment loads that likely accompanied settlement during the 19th century. The interpretation although reasonable does tend to be rather circular, however, because there are no sediment transport data for these rivers.

The same limitation applies to Schumm's (1968) interpretation of channel changes on the Murrumbidgee River in New South Wales, Australia. The patterns of change are evident in the landscape but the record of process largely is lost and must be reconstructed by means such as that offered by equations (4) to (20). Obviously such evidence cannot then be used to confirm the process-form relations!

Almost all major natural changes in channels occur so slowly that even if the morphologic record is well-documented, the process record is invariably inadequate. But there now may be an exception to this rule. Man, by his activities in and about rivers has created something of a laboratory of accelerated river change. For this reason Man's impact on rivers is of scientific as well as of immediate practical interest and has been the subject of much recent enquiry.

Human-induced Channel Changes

The study of the impact of human activity on river channels recently has been reviewed extensively by several authors (see Kerr, 1973; Gregory, 1976a, b, 1979, 1981; Park, 1977, 1981). For this reason it will best serve the present purpose

to simply consider some type examples of this research, noting general progress and future prospects.

Studies of human-induced channel change represent an extension of the empirical approach to river behaviour although there are very few actual detailed observations of channel change recorded. Those that are available are concerned either with the consequences of direct modifications such as dam construction, flow diversions, and channel stabilisation, or with the general but less direct effects of landuse change.

The Effects of Dams

By far the largest set of observations exists for the long-recognised downstream effects of dams, particularly the rapid channel degradation associated with the abstraction of sediment load. Many of these published studies recently have been reviewed by Makkavayev (1972) and Petts (1979). Dam closure typically results in substantially reduced flood peaks, increased base flows, and in the case of large reservoirs, almost complete abstraction of sediment load. General consideration of flow and sediment transport continuity (equations 1 and 3) suggests that this should lead to downstream channel contraction and degradation until bed armour or energy slope reduction stabilises the channel. The quality of these changes generally are confirmed by empirical studies (for example, see Lane, 1934; Shulits, 1934; Stanley, 1951; Livesay, 1963; Northrup, 1965; Wolman, 1967; Hammad, 1972; Dolan et al., 1974; Gregory and Park, 1974; and Rasid, 1979) although the degree of response is highly varied (for example, see Buma and Day, 1977, who describe a case of relatively insensitive response). Other effects such as changes in the regime of river ice and the rejuvenation of downstream tributaries (and consequent alluvial fan deposition in the main channel) are discussed by Kellerhals and Gill (1973).

Quantitative prediction of degradation below dams has been attempted by Tinney (1962), Komura and Simons (1967), Hales et al. (1970) and Komura (1971) but inevitably there is much discussion (see those by Herbertson et al., 1968) about the appropriate sediment transport equation to cast into differential form. Nevertheless, some success has been achieved in several cases in the United States although the lack of further testing leaves the general reliability of these semi-theoretical approaches unspecified.

Channel changes upstream of dams are less well documented but generally are recognised to involve reservoir infilling and aggradation upstream to the back-water limit.

Most empirical studies of degradation involve a simple before and after comparison of channel form generally over a period of 5 to 10 years. Far fewer studies involve ongoing regular monitoring of channel change although this is the type of data that clearly is needed. We know from the longer-term behaviour of rivers that the channel adjustment process commonly is complex and may take many decades to be achieved. A 5 to 10 year observation period may just be the introduction to a much longer and more complicated story in many cases. Experience with the American Vigil Network (Emmett, 1965, 1974; Leopold and Emmett, 1965, 1972) indicates that long-term monitoring of channel changes such as that currently being done on the Peace River below the Bennet Dam in British Columbia (by M. Church, pers. comm.) is an essential research strategy if the process of channel change ever is to be better understood.

Interbasin River Diversions

Similar comment might be offered, of course, in other cases of human-induced

channel change, including the converse of the dam closure situation: the inter-basin river diversion. The type of diversion of interest here is one in which a relatively large flow is diverted across a divide and allowed to flow into a dry valley or relatively small stream in which it develops a new larger river channel. Very little is known about the effects of such diversions because they are few in number (although often grand in scale), located in isolated and often uninhabited country, and generally are rather recent in construction. Kellerhals et al. (1979) have described eleven Canadian cases of interbasin river diversions and they may well represent the majority available anywhere for establishing useful precedents. They classify the diversions into three types: bedrock-controlled diversion routes, steep diversion routes in unconsolidated materials, and alluvial diversions. In the case of the first type (for example, the Churchill River diversion in Manitoba) channel changes are restricted to the flushing out of pockets of unconsolidated sediments and the erosion of weak bedrock. In the second type (for example, the Ogoki River diversion in Ontario) the diversion route is characterised by an expected rapid increase in channel capacity through channel widening and deepening accompanied by the export of large amounts of sediment (see Dzurisin, 1975, and Troxel, 1974 for an American example of this type). Alluvial diversions (for example, the Nechako-Kemano diversion in British Columbia) may remain little altered apart from moderate channel enlargement if the diverted flow is not significantly greater than the higher flood flows of the receiving stream. A most important conclusion of their study is that a detailed knowledge of all materials that might conceivably become exposed to erosion is an essential prerequisite to prediction of the final morphology of a diversion channel. This is particularly important, of course, in determining if and when bed armour will develop.

It is important to note, however, that most of the diversions considered by Kellerhals et al. (1979) are less than 20 years old. Yet even in the case of the 34 year old Ogoki River diversion, the receiving Little Jackfish River has not yet completely stabilised. Clearly there are many more lessons to be learned from their continued surveillance.

Channelisation

This term refers to the modifications made to channels in order to improve their navigability, lessen flooding, improve land drainage, or to make them more stable. Alteration of channels is common, to some extent, in all countries but in the United States it has been practiced on a massive scale involving over 13,000 river kilometres of levees and floodwalls and a similar additional length of channel "improvement". Many of these projects are summarised and analysed in reports for the Council on Environmental Quality (1973) and the Committee on Government Operations (1973). Although many channelisation projects clearly are successful (for example, see Canterbury, 1972) it unfortunately also is the case that many of the "improvements" involved changes to channel forms that we now know are unstable. For example, many rivers were straightened and thus steepened, causing downstream channel aggradation, bank failure and greatly increased sediment transport rates that exported the problems further downstream (for example, see Daniels, 1960; Ruhe, 1970; Emerson, 1971; Yearke, 1971; Swicegood and Kriz, 1973; Keller, 1975).

The classic example of this type of problem is the Mississippi River, probably the most manipulated large river in the world. The Middle and Lower Mississippi River provide an instructive contrast in the effectiveness of differing management strategies.

Maher (1963) and Stevens et al. (1975) describe some of the man-induced changes in the Middle Mississippi (between St. Louis and Cairo) that so dominate the

present channel character. The Corps of Engineers are charged with developing and maintaining a 2.7 m deep and 90 m wide navigation channel while at the same time providing flood protection. As a result, most of the Middle Mississippi is lined with Corps of Engineers mainline levees and much of the channel has been contracted to about two thirds of its former (1888) width by 150 km of diking system. The channel alignment is maintained by 200 km of bank revetment. It is possible that this engineering scheme simply hastened nature's work; the channel of 1888 was likely overdue, reflecting the effects of earlier extreme floods. In any event the minimum navigation channel largely is being maintained now and the flood hazard has been markedly reduced.

A similar engineering program was developed by the Corps of Engineers for the Lower Mississippi River (downstream of Cairo) but with one important difference: the channel was straightened by artificial cutoffs (see Winkley, 1972, 1973, 1977).

The cutoff program was a controversial one. On the one hand it seemed to be a means of improving the navigation channel and of reducing flood stages while on the other there were fears that local increases in slope would cause instability (see Humphreys and Abbot, 1859, for early support of this latter position). The outcome of the debate was that the cutoff program was approved in 1929.

The present natural length of the Lower Mississippi River, about 1740 km, appears to have been stable for the last 1000 years (Fisk, 1944, 1952), fluctuating between 1830 km and 1680 km in response to a natural cutoff frequency of about 14 per 100 years. Natural relaxation times following a cutoff seem to be 30 to 80 years (Winkley, 1977). Between 1929 and 1942 the artificial cutoff rate was increased to eight times the natural rate. In consequence, the river length was shortened to about 1530 km, mean water-surface slope was increased by 12 percent, while slopes locally were increased up to twenty times the pre-cutoff magnitudes.

The integrated effect of the downstream migration of the sediment slugs produced by each of these cutoffs has been channel widening, bar formation, and increasing bank instability in the southern portion of the Lower Mississippi (i.e. it is tending to braid). In spite of an accelerated maintenance program depths in the navigation channel are less than they were 90 years ago. Winkley (1977) estimates that the cutoff channels will not stabilise until the end of the century. The general aggradation problems presumably will persist well into the 21st century.

The effects of landuse

It has long been recognised that the character of the flood hydrograph is sensitive to landuse within the originating catchment (see the reviews in Chow, 1964). For example, the tendency for reductions in infiltration capacity, for whatever reason, to reduce lag times and increase flood peaks is one which is incorporated into most general flood forecasting models (the Stanford Watershed model provides a well-known example; see Crawford and Linsley, 1966). A considerable literature of hydrologic case studies designed to confirm or better define these relationships among landuse and flood characteristics has accumulated during the last few decades and many of these are discussed in recent general accounts such as those by Gregory and Walling (1973) and Dunne and Leopold (1978).

It also has been known for some time that basin sediment yields are strongly influenced by landuse (see U.S.D.A., 1964). Indeed, a distinction is made between "geologically normal" and the usually landuse-related "accelerated" rates of hill-slope erosion which can catastrophically increase the sediment supply to rivers.

Because channel size and form largely is the product of sediment supply and the annual flood, physical reasoning dictates that landuse change will promote channel change. Perhaps because it is both obvious and simple, this deduction rarely has been systematically tested although it has been confirmed often enough by casual observation. It is an unfortunate oversight and a potentially rewarding research area because the effects of landuse change on river morphology may provide a useful analogue of the response of channels to natural changes in the general environment.

The few studies of landuse-related channel change that are available are of two types. They either describe the history of channel adjustment to changing landuse or they describe the spatial variation in supposedly equilibrium landuse/channel morphology relations. Because of the difficulty of recognising such an equilibrium and of controlling for other potentially confounding conditions (for example, see Church and Mark, 1980 for a discussion of scale problems in geomorphology) the former approach at first may appear to be preferable. Direct monitoring of a site history, however, has two distinct disadvantages. First is the time constraint common to all direct monitoring and previously discussed in the context of meandering. Second, study areas are necessarily small and channel responses may be strongly influenced by site specific characteristics that make generalisation difficult.

Channel responses to two types of landuse conversion have received most attention because the changes involved are widespread and extreme and therefore more easily measured: those related to deforestation and those related to urbanisation. Although there is a small literature on the particular effects of mining on rivers (for example, see Bull and Scott, 1974; Graf, 1979; Park, 1979) they will not be included in the present discussion. However, the general discussion of channel response to changes in discharge and sediment supply to follow is directly applicable to cases of mining in or near river channels.

The impact of forestry practices on hydrology and channel morphology is not well understood in detail although there now are many case studies (of the historical type) described in the literature (see Penman, 1963; Sopper and Lull, 1967; Jeffrey, 1970; Bell et al., 1974; Janda et al., 1975). Perhaps few areas display such extreme responses to deforestation as are commonly observed in the Pacific Northwest of North America. Logging practices in Oregon, Washington, and British Columbia often involve clearcutting of the coastal forest on steep slopes of weak volcanics and glacial debris. Usually within a year of logging, slope failures abound, runoff and sediment yields markedly increase and much of the pre-logging drainage pattern on the upper slopes is obliterated by mass movement and devastating debris torrents (see Rothacker, 1970; Harr et al., 1975). The surviving channels are choked with huge volumes of sediment and logging slash, an important source of which is the spoil from logging road construction (see Anderson, 1970; Fredriksen, 1970; Burns, 1972; Megahan, 1972; Swanson et al., 1975). Replanting of these logged areas, however, will stabilise slopes and reduce erosion, sediment yield and runoff within a few years as the regrowth becomes established (for example, see Berndt and Swank, 1970).

The effects of partial vegetation removal on river flow in less extreme geomorphic environments is not well defined with studies often yielding rather different results for different areas (see Hibbert's inconclusive 1967 review of 39 case studies). The results of more recent studies are just as variable, however, and point to the prevailing sensitivity of channels to local conditions. Although the balance of studies seem to favour the notion that vegetation removal will increase water yields to some (unpredictable) degree, the effect on flood peaks

varies widely (for example, Harper, 1969, and Brown, 1972, find they increase; Hewlett and Helvey, 1970, and Lynch and Sopper, 1970, find the response negligible or rather variable; while Chang et al., 1975, find that they decrease!). Recent work stresses that, in small catchment studies, factors such as the area of logging roads, compaction of road surfaces, treatment of slash, variation in type and timing of storms, seasonality, etc., can dominate the effects of even extensive vegetation removal (see Ziemer, 1981). It is hardly surprising that comparisons among basins lead to inconclusive results. There is general agreement, however, that sediment yield is strongly dependent on the degree of disturbance to the soil (by skidding, hauling, road construction, etc.) and not on the removal of trees per se (see U.S.E.P.A., 1975; Rothwell, 1978). For this reason vegetation removal at one site may cause channel enlargement and incision (Carson and Tamm, 1977) while at another it will cause aggradation (Orme and Bailey, 1970, 1981; Janda et al., 1975).

There are no comparative (ergodic) studies of channel response to vegetation changes. Spatial variation in vegetation typically is associated with geologic, geomorphic, edaphic and climatic changes and it is not possible to isolate the precise influence of one from the others. In any case, adequate information on channel geometry, as noted earlier, simply is not available for analysis.

There are, however, a large number of U.S.D.A. studies on runoff and sediment yield for various agricultural and grazing intensities in the United States; these observations form the basis of the Universal Soil Loss Equation (F.A.O., 1965). In detail they show great variation in the degree of change in runoff and sediment yield from one landuse type to another but the direction of change is well established. Runoff and sediment yields are high from cultivated and heavily grazed rangeland (for example, see respectively, Ursic and Dendy, 1965; Noble, 1965) and relatively low for forests and ungrazed rangeland; all yields increase with increases in precipitation.

Langbein and Schumm (1958) and Rango (1970) analysed sediment yields on a regional basis and concluded that as precipitation increases, they increase rapidly from zero to a temperature dependent maximum and subsequently decline because of the corresponding increase in the protective vegetation cover.

Although these runoff and sediment-yield data are rather variable and inadequate, for precision modelling, together they do suggest qualitative hydrologic and sedimentologic changes associated with vegetation change. Unfortunately, however, channel changes are extremely sensitive to the balance between discharge and sediment supply and these data often do not permit the balance or the quality of channel changes to be specified.

An interesting case of direct modification of channels by vegetation is described by Graf (1978). He used historical photographs to document a 27 percent reduction in the width of the Green River (Utah) that he ascribed to the spread of tamarisk in the region. Although his argument is somewhat weakened by the fact that mean annual discharge also has declined by about 25 percent in the same period, there is little doubt that channel morphology is influenced by changes in bank vegetation.

The effects of urbanisation on rivers also have been the subject of intensive investigation over the last few decades but once again far more is known about urban hydrology (see Leopold, 1968; Douglas, 1976) than is known about the morphological response of rivers in urban areas (Graf, 1975, 1976, 1977; Cooke, 1976).

The increase in flood peaks and decrease in lag times associated with the

spread of impervious surfaces and development of efficient sewerage systems in urbanising areas have been recognised for many years (for example, see Savini and Kammerer, 1961; Van Sickle, 1962; Anderson, 1963). More recent historical studies of urban development have confirmed and consolidated these earlier findings (for example, see Crippen, 1965; Kinoshita and Sonda, 1969; Seaburn, 1969; Walling and Gregory, 1970; Gregory, 1974; Hollis, 1974, 1975; Schneider, 1975). Further support comes from a large number of comparative studies of paired rural and urban watersheds (for example, see Stall and Smith, 1961; Ikuse et al., 1975; Lai, 1980) and from studies combining both approaches (Waananen, 1969). These observations have provided the basis for simple empirical models for flood-peak prediction (Carter, 1961; Espey et al., 1966; Demster, 1974) and for general urban hydrology models such as the Stormwater Management Model in the United States and the Road Research Laboratory Hydrograph Model in Britain (see Larson, 1972; Roesner et al., 1972; Aitken, 1973).

In general these studies show that urbanisation typically will double flood peaks and halve lag times although Espey and Winslow (1974) and Hollis (1975) note that such changes depend on the recurrence interval in question (for example, the relative effects appear to decline as the recurrence interval increases). They obviously are also dependent on the character of urbanisation (area of roads and parking lots, detached or integrated housing, etc.; see Dunne and Leopold, 1978) and on the efficiency of the designed drainage system (see Beard and Chang, 1979).

In contrast to the case of logging, the effect of urbanisation on flood peaks is relatively direct and straightforward. The impact of urbanisation on sediment yields, however, appears to be very similar to, but far less extreme than, that observed in logged areas where typically sediment is initially mobilised in great amounts and then yields decline as secondary vegetation becomes established. In the urbanisation case the initial construction period commonly is characterised by increases in sediment concentration and yield by respectively one and two orders of magnitude (Keller, 1962; Wolman and Schick, 1967; Walling and Gregory, 1970; Guy and Jones, 1972; Walling, 1974; Wolman, 1975). After the construction phase sediment yields appear to decline with the decreasing availability of sediment sources to levels possibly as low as those associated with forested basins (Wolman, 1967a, b; Wolman and Schick, 1967).

As before, physical reasoning might lead us to deduce that urbanisation initially will result in heavy sedimentation and contraction of stream channels and in subsequent channel enlargement to accommodate the new regime of reduced sediment yields and increased flood peaks. Indeed, there is some empirical evidence to support this sequence of events (Wolman, 1967a, Graf, 1976). The historical study of Watts Branch in Maryland (Leopold and Emmett, 19-2) shows that urban encroachment on to the headwaters there has caused a progressive decrease in channel size in response to increased suspended sediment loads since observations began in 1953 (Emmett, 1974b). The rate of channel contraction markedly increased after 1960 (Leopold, 1973) until, by 1972, the channel had been reduced to about 80 percent of its 1953 capacity. The frequency of flooding has doubled and Watts Branch presumably remains in a transient phase of adjustment to the altered hydrologic regime. Unfortunately this case study appears to be one of a kind. York and Herb (1978) describe a similar study in the Rock Creek and Anacostia River Basins in Maryland but only two surveys of channel morphology were made and the use of sediment control structures makes the results difficult to compare with those obtained for Watts Branch.

There is other evidence available, however, in the form of comparative studies

of urban versus "natural" channels. The relation of the enlargement ratio urban/natural channel size (scaled by drainage area) to landuse was first systematically examined by Hammer (1971, 1972) for 78 small watersheds near Philadelphia. He found that the enlargement ratio was high for areas with sewered streets and for those with major impervious areas such as parking lots, and that it was considerably smaller for areas with unsewered streets and for those impervious areas with detached houses. His data suggest that there is at least a 4-year lag between urban development and channel response and that channels in areas of streets and detached housing curiously have contracted to their former pre-urban dimensions by the time they exceed 30 years in age. Hammer (1972) notes that the enlargement ratios accord with Leopold's (1968) average annual flood ratios for corresponding landuse types.

Similar studies have been repeated in England by Hollis and Luskett (1976), Gregory and Park (1976) and Park (1977) with similar but somewhat more variable results than those reported by Hammer (1972).

One of the important facts emerging from this type of study of urban channel enlargement is that the results of comparative analyses include considerable "noise" related to variations in lag time (response and relaxation times). Lag times likely are dependent on both the scale of the system (drainage area) and on stream power (for example, Hammer's, 1971, data indicate that channel slope is an important statistical determinant of the amount of channel enlargement) among other things. But we simply do not know at present. This, of course, is yet another compelling argument for the need of further historical studies such as that on Watts Branch. It also is important that observations not be restricted to a single reach; degradation in one reach presumably implies aggradation somewhere downstream. To this end of characterising changes within the entire drainage network, Gregory (1977) suggests that changes in the channel network volume might better reflect the integrated channel response (see also Gregory and Ovenden, 1979).

Channel Changes in Geomorphic Time

It is not the purpose of this section to review the vast array of case studies describing rivers displaying evidence of long, varied, and unique histories. Instead, it will very briefly consider some of the problems and prospects of identifying and interpreting channel changes that have taken place prior to the period of recorded observations.

The primary evidence of channel changes in geomorphic time is the morphology and stratigraphy of alluvial valley fills. Detailed description of this evidence remains essential to any analysis of fluvial history although with the recent emphasis on contemporary process studies, such detailed field descriptions seem to be less common now than they were in the past. Three well-known examples, amongst many others, characterise this type of study: Fisk's Holocene history of the Mississippi Valley, Dury's study of regional underfitness of rivers, and Leopold and Miller's study of the alluvial chronology of valleys in the American southwest. Fisk (1944, 1947, 1952) described some 15,000 years of Mississippi River aggradation based on a knowledge of the detailed stratigraphy of the Mississippi Valley set in the general context of the late Quaternary climate and sea levels established elsewhere (see Flint, 1971). His reading of the stratigraphic record suggested to him an early deglacial of very high sediment supply in braided channels (extensive deposits of sands and gravels) giving way to a meandering Mississippi as sediment supply, sediment calibre and valley slopes declined to the present. Schumm (1968) cites Fisk's work as a confirmation of the concept of river metamorphosis. Dury (1964a; see his brief review, 1977) identified and surveyed former large channels buried by Holocene alluviation in England, France, Wisconsin and Australia in order

to test his theory that misfit streams are widespread and of climatic origin. He used drilling and seismic surveys to delineate the bed profiles of buried rock-cut channels and stratigraphic, palynological and carbon dating evidence to reconstruct the infilling chronology from the time of channel abandonment some 9,000 - 12,000 years B.P. He went on to use general hydraulic geometry and regional climatic norms to retrodict Holocene hydrologic and climatic conditions. Leopold and Miller (1954) and others (see Hadley, 1960) surveyed valley fills in Wyoming and using stratigraphic, pedologic, carbon dating, fossil and archaeological evidence, developed chronologies involving major alternating phases of degradation and aggradation since late Pleistocene times. Although the general chronology neither is complete nor entirely synchronous at all sites, an idealised valley schema involves three terraces in three sedimentary units (see Haynes, 1965). The oldest dates from late Pleistocene (fossil evidence) and was formed by channel incision during an arid phase (the upper terrace sediments include dune sands and calcic palaeosols). Further alluviation and subsequent incision produced another terrace about 600 - 800 years B.P. in similar climatic circumstances. A final major period of deposition followed and ended with the initiation of modern gullying in 1880-1890 and the formation of the most recent terrace. The reason for the most recent period of gullying is unclear but likely reflects general climatic shifts and in some places overgrazing of domestic animals. Vigil Network observations suggest that many of these valleys have been in a depositional phase during the second half of this century (Leopold, Emmett, and Myrick, 1966; Leopold and Emmett, 1972) when the region generally has been experiencing a cooler and wetter climate. The record is by no means clear, however, because of the complicating effects of human activity (see Meade and Trimble, 1974; Knox, 1977) during this century.

Each of these studies shows how careful reading of the sedimentological record can yield two types of information about channel changes. The first is the actual chronology of change and the second is the process information implied by the observations. Amid the controversy that so often surrounds the interpretation of this process information, the value of the long-term record of channel behaviour per se sometimes seems to be overlooked. Only in this longer-term context can we develop a realistic perspective of present and future channel behaviour in the short term.

Much of the controversy surrounding the interpretation of alluvial chronology results from a certain circularity of argument. For example, in the development of alluvial chronology for the American southwest (and for the Riverine Plains in eastern Australia; see Butler, 1958; Schumm, 1968) there has been a tendency to evaluate channel changes in terms of a-priori models of climatic change and river response and then to argue for those assumptions from the field evidence. This problem, of course, simply points to the need for an independent climatic and hydrologic record against which channel changes can be evaluated.

In this context it seems that far too little attention has been paid to the potential of extracting such information from the vegetation record. In particular, recent developments in the analysis of palynological (Bryson and Kutzbach, 1974) and dendrochronological (Fritts, 1976; Shroder, 1980) records appear respectively to provide some access to Holocene and detailed access to the last few centuries of climatic and hydrologic conditions. Two recent examples from North America illustrate the potential of these developments.

Mathewes and Heusser (1981) present 12,000 years of mean annual precipitation and mean July temperature records implied by the column of fossil pollen sampled from a lake in southwest British Columbia. Transfer functions for converting pollen frequencies to climatic parameters were adopted from Heusser et al. (1980)

who sampled modern pollen data from 180 coastal sites from the Aleutian Islands to California. They related the frequency of four pollen factors to the modern (1968-1977) July temperature and annual precipitation records at 43 meteorological stations to yield a pair of regression equations used as transfer functions in the Mathewes and Heusser (1981) lake study. Eight radiocarbon dates and a Mazama ash layer provide chronostratigraphic control of the lake core. This proxy climatic record shows an approximately 2°C increase in mean July temperature rising from minimum values about 12,000 years B.P. to maximum values about 10,000 years ago. Temperature declines again to near late glacial levels by 6500 years B.P. and subsequently rises gradually by 0.5°C to the present. Precipitation declines from late glacial times to a minimum at 7000 - 8000 years B.P. and then rapidly increases to present levels by about 6000 years B.P. after which it is maintained to the present.

Other pollen/climate transfer functions have been developed elsewhere (see Parker and Henock, 1971; Webb and Bryson, 1972; Bryson and Kutzbach, 1974; Kay, 1979; Andrews et al., 1980) and as the data base increases this technique should provide increasingly more reliable estimates of Holocene climate.

In a methodologically similar study, Duvick and Blasing (1981) cored about 100 trees in central Iowa and determined a 300-year chronology of tree-ring width indices in accordance with procedures proposed by Fritts et al. (1971, 1979). They calibrated this record with statewide precipitation records for the period 1920 to 1979 and tested the resulting regression equation by predicting the actual record for the period 1874 to 1919. As a result, the tree-ring model was shown to be capable of explaining about 55 percent of the variance in the test record. This level of explanation makes the model a reliable indicator of wet and dry periods in central Iowa dating from the year 1680. Again, this is but one of a growing number of published proxy precipitation records (see Fritts, 1976) and their reliability and range presumably will increase as the data base expands.

Perhaps Nanson's (1980) study of the Beatton River in British Columbia is a good example of the approach that should be taken when examining river activity. He combined a survey of contemporary channel form and process, tree-ring analyses of lateral and vertical channel movement and detailed stratigraphic and sedimentological analyses, to develop a record of channel change for the last 400 years. In this case, as in so many others, the character of the present channel is dominated by processes that only display their full range of behaviour in geomorphic time.

Some Concluding Remarks

Some of the more important conclusions in the foregoing discussion can be summarised as follows:

1. If the concept of equilibrium is to be usefully applied in the context of channel changes, it must be realised that its meaning changes as our time frame widens and contracts. Although this is hardly a novel idea, perhaps only now, with the appearance of such accounts as those provided by Cullingford et al (1980), is it receiving the serious attention it deserves.
2. A valid process-based theory of equilibrium channel morphology at an engineering time scale does not exist. The process equations that would constitute such a theory are either poorly defined or undefined.
3. The process of sediment transport, central to all types of channel changes, is described very poorly by equations that may be theoretically unsound; all

assume, contrary to at least some experience, that sediment transport rates are not supply limited.

4. The empirical view embodied in regime "theory" and the concept of hydraulic geometry probably represents the only viable approach to prediction of channel changes. At the present time, however, there is a remarkable paucity of data. It is likely that the most useful data will be those collected for environmentally homogeneous regions.
5. There is an urgent need to monitor a variety of channel changes through long periods of time. Only from studies such as those presently constituting the U.S. Vigil Network will reliable data on rates of channel adjustment, including response and relaxation times, be obtained. These studies should include a wide range of fluvial system scales.

A theme woven into much of this discussion is that river morphology, particularly from the engineer's point of view, generally may be a non-equilibrium property. The literature is replete with examples of rivers for which the significant formative processes are those only viewed properly at a geomorphic timescale. Perhaps most rivers are dominated by transient behaviour, never fully adjusting to such events as major floods, climatic shifts, and those step function effects known as river metamorphosis. Church (1980) has placed this latter effect in the general context of the Hurst phenomenon, one cause of which are stationary stochastic processes whose parameters give rise to substantial low frequency events.

Perhaps it is time to heed the remarks of Burkham (1981) who, after assessing the ability of our fluvial science to respond to the information needs of environmental law, concludes that its dominating characteristic is uncertainty. It seems likely that an appreciation of the scope of this uncertainty will only come from studies of the long-term river record. One of the more significant gaps in our knowledge is the nature of the river record during the last several hundred years. This would seem to signal a new emphasis on the perspective and techniques of the Quaternary geologist. In the end, it may be that, indeed, uncertainty is the mark of our fluvial science; that would not be an insignificant conclusion.

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