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Spatial and temporal variations in erosion and sediment yield

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ABSTRACT A key issue in river basin studies concerns the temporal and spatial variability of erosion and the movement of sediment within the drainage basin. Much of the theory involving the relationships between these processes has been developed from studies of relatively small areas over short time spans. Extrapolation of this body of theory to larger areas over longer time scales has proved extremely difficult as the complexity and heterogeneity found at the larger temporal and spatial scale precludes synthesizing the sediment flux response from information obtained at much smaller scales. Increases in spatial scale involves increases in complexity, dimensions, new variables, new relationships, and as a rule leads to the identification of new problems. Extrapolation of short term observations over longer time periods involves frequently unjustified assumptions about the sequence, magnitude and frequency of erosion and sediment movement. A review of twenty-five years of study of erosion and sediment production in southern Alberta in basins ranging from 44 000 km^2, to a few thousand square metres, to small runoff plots demonstrates that it is essential to make observations and develop theories at the spatial and temporal scale compatible with the drainage basin of interest in order to avoid formulating unrealistic theoretical models.

INTRODUCTION

In his comments on the contributions of W. M. Davis to geomorphology Leighly (1940) stated that "Davis's great mistake was the assumption that we know the processes involved in the development of landforms. We don't, and until we do we shall be ignorant of the general course of their development." This assessment was followed in the 1940's and 1950's by a considerable effort to put geomorphology on a more quantitative, process-oriented basis. Building largely upon Horton's (1945) seminal paper on the relationship of erosional processes to overland flow, and Strahler's (1950; 1952) trend-setting morphometric approach, geomorphologists began to redefine the concepts of landform development using quantitatively-based models. By attempting to link spatial and temporal concepts in order to understand the nature of the variables that were involved (Schumm and Lichty, 1965), geomorphologists were faced with the dilemma of resolving short-term (seasonal, annual or at most decade-long), high resolution process-study data and observations with very much longer term (centuries or millenia), low resolution landform interpretations based on standard, geologically-based analytical techniques. The problem of satisfactorily integrating the results of these two sets of information remains an often intractable one and our general inability to understand and measure the relationships between the large range of processes involved in erosion and sediment yield is a prime example of this difficulty.

SPATIAL AND TEMPORAL HETEROGENEITY

One of the principal concerns across the wide range of disciplines dealing with river basin studies is the often enormous variation in the spatial and temporal patterns of erosion, sediment transport, sediment storage and sediment yield. The processes of erosion and sediment production within any basin, no matter what its size, are spatially and temporally discontinuous and the variance increases with increases in spatial and temporal scale. Studies which relate sediment yields to basin-wide parameters such as, for example, total area or regional climate and to use such data to show or predict patterns of erosion at any scale - local or global - are a case in point (e.g. Fournier, 1960; Menard, 1961; Corbel, 1964; Judson and Ritter, 1964; Strahkov, 1967; Holeman, 1968; Walling and Webb, 1987; Jansson, 1988). These studies, and the generally very wide variations in sediment yields and regional erosion rates that they portray, illustrate the problem of extrapolating highly localized, short-term data over long distances of time and space based on information which is almost always inadequate and frequently inaccurate. In her extensive review of global climatic variations and their influence on sediment yield, Jansson (1988) concluded that there is no simple relationship between runoff and sediment yield. There is no reason to suppose there would be. To paraphrase Philip's (1975), comments on basin hydrology, both sediment yields and runoff are the aggregation of a spatially and temporally heterogeneous system in which, because of the variation of the parameters within the system, any one of a vast number of variations may yield the same output. As Haggett et al. (1965) point out, increases in spatial scale involve increases in complexity and spatial dimensions, new variables, new relationships and as a rule lead to the identification of new problems. And the same comment can be made with reference to temporal scale; the result is that we have little understanding of the variance across different scale levels (Thornes and Brunsden, 1977).

Temporal variations in factors such as time lags between erosion, sediment movement, storage and remobilization also create difficulties in predicting historic patterns of sediment yields based upon modern erosion rates (Trimble, 1975, 1977). Trimble (1983) shows not only the spatial variability within a fairly homogeneous area, he demonstrates that sediment yields between 1853-1938 in the Coon Creek basin, Wisconsin, were only c. 6% of the erosion that was estimated to have occurred. As Church (1980) observes, assumptions about temporally-based interpretations of geomorphic events may not be justified without detailed knowledge of the timescale of significant landscape changes.

Drainage basins probably only rarely conform to the model of an upper basin zone of sediment production, a middle zone of sediment transport and a lower basin zone of sediment deposition (Schumm, 1977; Sundborg and Rapp, 1986; Starkel, 1991). Only in the extremely long span of geological time is the sediment yield ratio likely to approach unity - a proposition which is at least implied in the basin tripartite model - and even in this time frame it is unlikely that the basin will have the landform stability and time needed to complete such an erosion-sediment yield cycle.

Drainage basins are fuzzy systems (Fig. 1). The patterns of runoff and sediment flux are dynamic and ever fluctuating. While the absolute, hard boundaries of the basin system can usually be determined - divides and channel beds - even these change over time: channel bed configuration, due to often rapid variations in deposition, storage and erosion of sediments, is a particularly active boundary and, given the difficulties of measuring bedload, can be a major problem in estimations of sediment storage and yields.

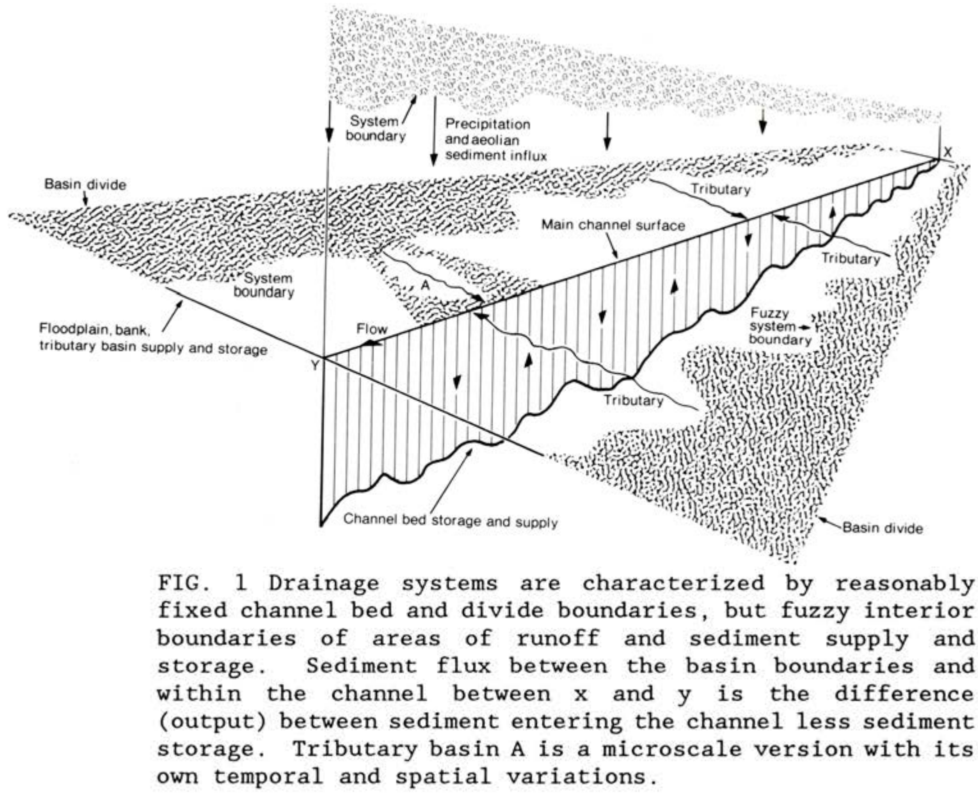

FIG. 1 Drainage systems are characterized by reasonably fixed channel bed and divide boundaries, but fuzzy interior boundaries of areas of runoff and sediment supply and storage. Sediment flux between the basin boundaries and within the channel between x and y is the difference (output) between sediment entering the channel less sediment storage. Tributary basin A is a microscale version with its own temporal and spatial variations.

Within the system boundaries, fluctuations in the partial area contributions of runoff and sediment (Campbell, 1985) mean that each, individual, precipitation event involves a different, heterogeneous mix of sediment and runoff producing areas. The intermittent nature of sediment movement in the basin, with its often lengthy storage times, makes the establishment of accurate sediment budgets extremely difficult unless the storage sites can be identified, their sediment volumes determined, and the periodicity of sediment flux accurately dated (Baker, 1983).

The movement of sediment from the eroding site in the basin may take millenia to appear at the basin mouth. While the flow of solute load closely mirrors the rates of stream flow, the rates of movement of suspended load and bedload become increasingly slower so there is not even a general synchronicity of the various phases of sediment movement.

The mean velocity of the total instream sediment transport may be barely perceptible (Campbell, 1989).

The result is that in many studies there is often an inappropriate mix of temporal and spatial scales of investigation. Microscale temporal and spatial studies involving detailed analysis of modern processes show little if anything about the meso and macroscale evolution of the watershed (Church, 1980). The reactivation of sediment from a long-term, unidentified storage site in the channel bed, or from the floodplain banks adjacent to the channel, supplies sediment to the stream which, when measured at a gauging station, is frequently interpreted as representing erosion from some upland source within the basin. Hence the wide variations in calculated regional erosion rates and sediment yields.

Yet, the need to transfer spatial and temporal data across various scales is fundamental to geomorphology, so that it is critical to know the range of scales over which data transference may be validly made. Examples taken from southern Alberta can be used to illustrate the difficulty of relating patterns of erosion and sediment yields across a wide range of temporal and spatial scales.

EROSION AND SEDIMENT YIELD, RED DEER RIVER BASIN

The Red Deer River basin in Alberta has a gross drainage area of 44 682 km^2 (PFRA, 1989) above the lowest downstream gauging station at Bindloss (Fig. 2). The 'gross drainage area' is defined (PFRA, 1983) as the planimetric area enclosed by the drainage divide which should contribute runoff during extremely wet conditions. Recognizing the theoretical limitations of those conditions the PFRA (1983) also identified the 'effective - or dry - drainage area' as that portion of the drainage basin which might contribute runoff during a flood with a return period of two years. Thus, the 'effective area' of the Red Deer basin above Bindloss is reduced to 31 653 km^2; or about 70.8% of its gross area. Almost all of this reduction occurs in the semi-arid prairie portion of the basin (Fig. 2) where because of low precipitation and high evaporative losses, runoff coefficients are typically very low. Underhill (1962) calculated that the prairie portion of the Red Deer basin would generally have annual runoff values of about 50 mm near the city of Red Deer, about 10 mm near Drumheller, and only about 1 mm near Bindloss. While these values may be true for the flat, prairie surface, they are not true for some parts of the Red Deer river valley.

About 800 km^2 of steep, barren badlands terrain lines the Red Deer for about 300 km from just upstream of Bindloss to a point about midway between Drumheller and Red Deer (Fig. 2). The character of that landscape favors large, rapid runoff, and huge amounts of sediment. Comprehensive and detailed studies of badlands erosion processes at a large range of spatial and temporal scales have been conducted over the last twenty-five years. These studies reveal the difficulties of attempting to explain the patterns of erosion and sediment yield within the Red Deer basin in the absence of an understanding of the spatial and temporal heterogeneity of the landscape.

The mean annual sediment load of the Red Deer river at Bindloss is about 1.92×10^6 t (Fig. 3). At Drumheller it is of the order of 530×10^4 t, and at Red Deer it is about 199×10^4 t (Fig. 3). Between Drumheller and Bindloss then, almost 1.5×10^6 t of sediment is apparently added to the Red Deer. This massive injection of sediment is derived almost entirely from the 800 km^2 area of badlands. Thus, less

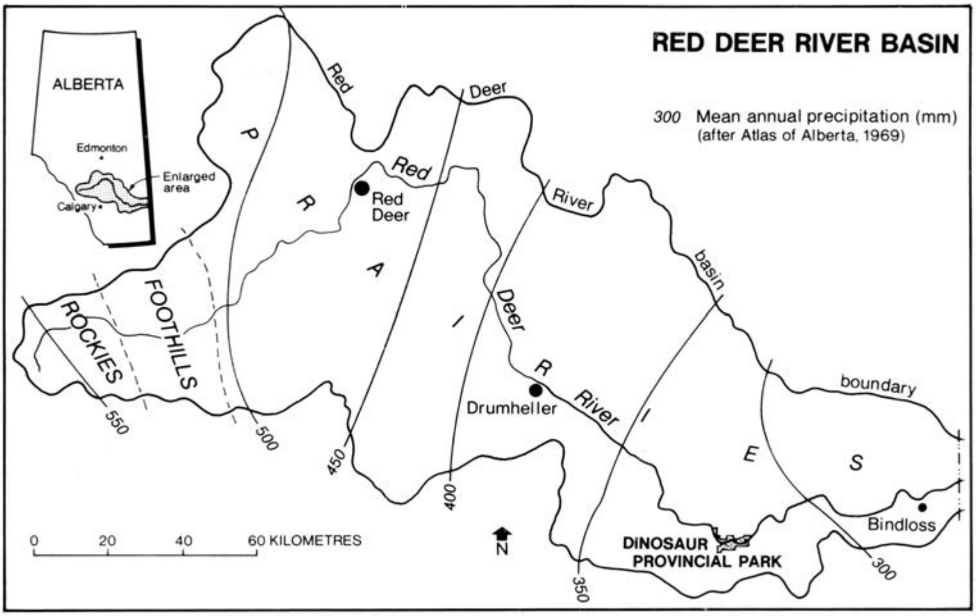

FIG. 2 The Red Deer basin (c. 44 000 km²) has suspended sediment sampling stations at Red Deer, Drumheller and Bindloss. Most of the sediment in the river downstream of Red Deer is produced from rapidly eroding badlands that line the river between Red Deer and Bindloss. The badlands (800 km²) form less than 2% of the area of the basin but contribute over 70% of the sediment load of the river.

than 2% of the gross area of the Red Deer basin potentially contributes over 70% of the total load. In fact this percentage is higher because the badlands upstream of Drumheller undoubtedly contribute most of the load at that gauge (Campbell, 1977).

Therefore, at the macroscale spatial level a microscale sediment source area apparently dominates the pattern of sediment yield within the Red Deer basin. The temporal pattern of erosion is less clear.

The badlands formed shortly after the retreat of late Wisconsinan ice from southern Alberta c. 14 000 years ago (Bryan et al., 1987) following the incision of much of the present Red Deer valley as a glacial meltwater channel. Rapid erosion in the weak, Upper Cretaceous shales which form the regional bedrock favored the formation of badlands (Campbell, 1987).

The 800 km² area of badlands can be viewed as a system of giant gully-like forms etched into the Red Deer's valley walls. On average, each gully is a triangular-shaped erosional wedge about 100 m deep and approximately 500 m long. About 20 km³ of material were removed largely by immediate post-glacial erosion.

The present Red Deer floodplain is close to 1 km wide along the 300 km long stretch of badlands and, although the depth of alluvial fill undoubtedly varies, the best estimate (based on bore holes) is that at least 20 m of alluvium fills the present valley (McPherson, 1968). Thus, some 6 km³ of alluvium - as a minimum value - is presently in storage in the badlands section of the Red Deer valley. But there is little evidence to indicate the age-range of those deposits; the basal units may be thousands of years old. The height of the present floodplain (lower terrace?) varies from 8-10 m above the low flow level of the river and much of this area has been inundated in historic times. The installation of the Dickson Dam upstream of the city of Red Deer has, since the mid-1980s, had a marked effect in reducing both flooding and sediment transport (Fig. 3).

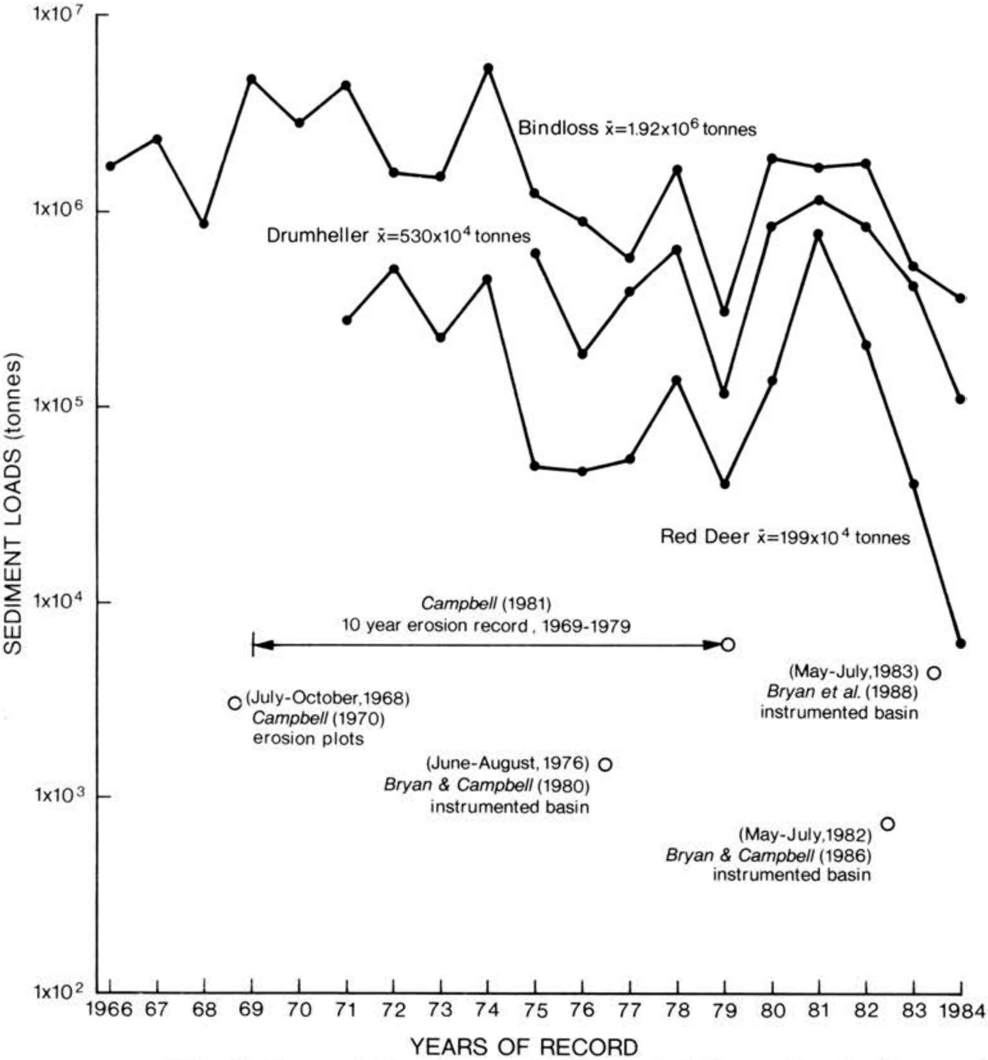

FIG. 3 Suspended sediment loads at Red Deer, Drumheller and Bindloss together with sediment yields per equivalent 1 km^2 area calculated from referenced field studies.

A sixty-five kilometre long section of the Red Deer valley (Fig. 4) shows the general relationship of the valley to its flanking badlands in the vicinity of Dinosaur Provincial Park. The valley is about 1 km wide and constrains the meandering Red Deer within the 100 m high walls of the glacial meltwater channel. A regular pattern of point bars characterizes the valley floor.

Detailed, largely independent field experiments and erosion and sediment transport monitoring studies (Campbell, 1970; Bryan and Campbell, 1980; Campbell, 1981; Bryan and Campbell, 1982; 1986; Bryan et al., 1988), confirm that mean annual erosion rates in the badlands of Dinosaur Provincial Park vary between c. 2.0-4.0 mm $year^{-1}$. A decade-long erosion grid study (Campbell, 1981) indicated somewhat higher rates of erosion but this study involved continual monitoring while the other studies were summer-only in duration. Expressed in terms of sediment yields per 1 km^2 area (Fig. 3) the studies indicate that on average about 2500 t $year^{-1}$ of sediment are eroded from the badlands. Because of the steep nature of the topography, and the generally extremely restricted possibilities for sediment storage in the badlands, almost all of the eroded material is delivered either into the Red Deer via tributary streams or stored as slope and channel wash deposits on the Red Deer floodplain (O'Hara, 1986).

Given the 800 km^2 total area of badlands, and the generally representative value of 2500 t $year^{-1}$ of badland-produced sediments there is a potential supply of c. 2 x 10^6 t of sediment to the Red Deer; a value very close to the average suspended load measured at Bindloss (Fig. 3). What is not possible to determine, however, is whether the load at Bindloss represents modern badland erosion, re-working of the older valley fill or, what is more likely, some indeterminable combination of both.

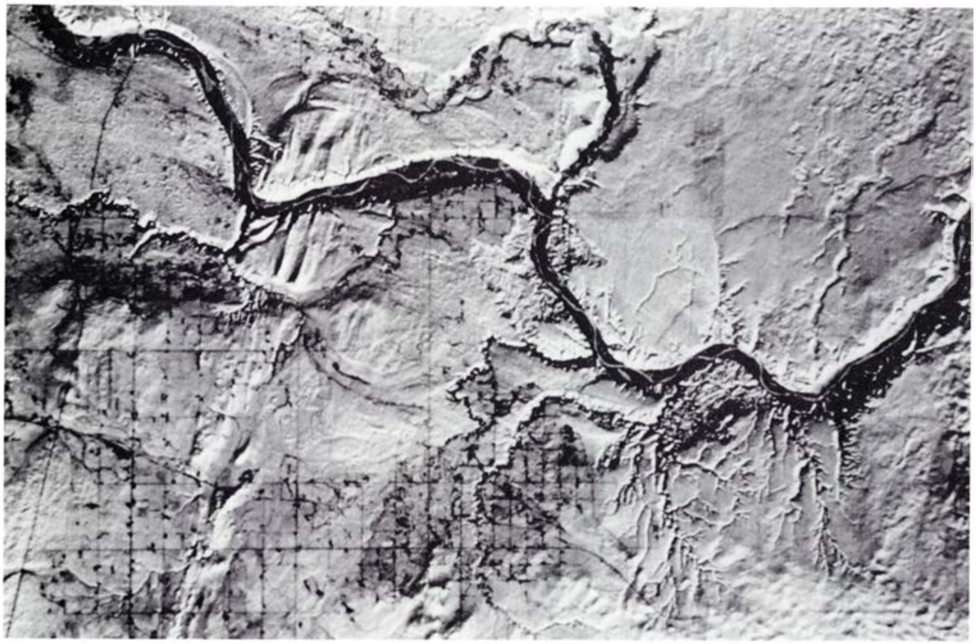

FIG. 4 Landsat 5 image of lower portion of Red Deer river near Dinosaur Provincial Park (badlands area in southeast). Image measures about 47 km east-west and 32 km north-south. Extensive point bars line the 100 m deep glacial meltwater channel which constrains the river.

The presently active and extensive area of point bars which line the valley of the Red Deer (Fig. 4) represent a large store of potential sediment supply to the river. Aerial photographic interpretation spanning a period of eleven years (Kondla and Crawford, 1971) indicated a mean erosion rate of c. 3.5 m $year^{-1}$ on point bars in Dinosaur Provincial Park, and Neill (1965) found a similar down valley migration rate of point bars near the Duchess bridge, just up-river from the park. The average size of the point bars is about two kilometres long and close to 500 m at their widest point, and they average nearly 8 m in height. So there is a potential sediment supply of some 20-30 000 m^3 $year^{-1}$ from each point bar; this would equate to a mass of about 40-60 000 t $year^{-1}$ from just one point bar.

Such a rate of erosion along the 65 km long stretch of river shown on Fig. 4 would produce a value very close to the mean annual sediment load of 1.92×10^6 t $year^{-1}$ recorded at Bindloss. And the erosional loss on the point bars would be balanced almost exactly by the input of sediment eroded from the badlands.

CONCLUSIONS

At the very small temporal and spatial scales of runoff plot studies (Campbell, 1970) it is a simple matter because of the high degree of resolution, to identify the highly erodible surfaces and to account for the particular patterns of sediment yield that are associated with specific runoff events: the data are both spatially and temporally, compatible and the relationships between the various processes are relatively easy to understand.

But, as the temporal scale increases from a summer-only study (Campbell, 1970) to a decade-long study - while the spatial scale remains at the same micro-level (Campbell, 1981) - the time variance in the behavior of the runoff plots becomes much larger than that due to differences in the erodibility variations of the individual plots (Campbell and Honsaker, 1982). Thus, the spatial and temporal relationships become blurred and the micro thresholds that are identifiable when both spatial and temporal scales are very small are unobservable (de Boer and Campbell, 1989).

When the spatial scale increases to the meso-basin size of several thousand square metres (Bryan and Campbell, 1980; 1982; 1986), but the temporal scale remains restricted to summer-only observations, the ability to identify the location and effects of high-sediment yielding surfaces is lost. It is the location, size and intensity of runoff inducing precipitation patterns and antecedent moisture conditions within the basin that produce the major variations in sediment yield (Bryan and Campbell, 1986).

At the very large, macroscale level of major drainage basins, like the Red Deer, the use of microscale temporal studies based on a few samples of stream sediment loads to determine the sediment yield pattern of the basin is an inappropriate mix of spatial and temporal scales. Even in the case of the Red Deer, where the high sediment yielding surfaces are readily identifiable, it is not possible, despite extensive investigation, to determine the contribution of present erosion on the badlands surfaces to the current sediment load of the river.

In most of the world's drainage basins the principal sediment source areas have not been identified and there is no accurate base on which to ascertain either from where or when the sediment load is being derived.

While our knowledge of the processes responsible for producing sediment has, at least at the microscale spatial and temporal level, advanced significantly following Leighly's (1940) admonition, our grasp of large scale variations in spatial and temporal relationships remains fuzzy. The greater the degree of difference in spatial and temporal scale the less valid become attempts to generalize from one set of observations to another and the mixing of scales of observation produces an even greater level of uncertainty (de Boer and Campbell, 1989). Unless we use scale-compatible models we shall continue, as Philip (1985) observed, "performing irrelevant calculations on an erroneous model of the real world".

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