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FLOOD-RELATED CHANNEL CHANGE IN AN ARID-REGION RIVER

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ABSTRACT

A review of 112 years of change in the channel of the Salt River, central Arizona, U.S.A., shows that this arid-region river has a main-flow channel that has migrated laterally up to 1·6 km (1 mi) in response to floor events. Maps showing locational probabilities indicate that along the channel zones of relative locational stability alternate with zones of relative instability at a 3·2 km (2 mi) interval. Construction of upstream reservoirs has reduced sediment input into the main river but has not controlled floods. The channel width has not changed except for moderate fluctuations around mean values; the main-flow channel has incised approximately 6 m (20 ft) over most of the 48 km (30 mi) study reach during six recent floods. Gradient has remained unchanged. During floods bed material was mobilized to a depth below the original bed level that was greater than the height of the water surface above the original bed. Calculations based on tractive force indicate a threshold discharge of instability that is equal to the flow with a five-year return interval. The river exhibits remarkable stability with respect to gradient and sinuosity, irrespective of water and sediment discharges, but horizontal channel location exhibited selective instability. Over the record period of more than a century, the channel appears not to have been in equilibrium considering geometry, discharge, and sediment.

KEY WORDS Channel change Arizona Hydraulic geometry

INTRODUCTION

Many of the basic principles of modern fluvial geomorphology depend on assumptions of a general system operating on a continuous basis and tending toward some equilibrium state. From the time of Dana (1850) and Gilbert (1877) until relatively recently, these assumptions provided great advancements in understanding of stream processes and have spawned positive working relationships with hydraulic engineers who emphasized the concepts of regime theory (Shen, 1979). As recent workers have begun investigation of an increasingly wide range of environments, however, the established theories based on continuous operation have not been found to be universally applicable, a trend common to fields other than geomorphology (Prigogine, 1978).

Rivers in arid regions pose difficult, if not impossible, applications of long-term equilibrium concepts because arid fluvial systems are markedly discontinuous in their operations and because the ratio of flood discharges to average discharges is so large (Cooke and Warren, 1973, review arid fluvial systems; see also Baker, 1977). Most geomorphic work in these systems, at least so far as changes in channel configuration are concerned, takes place during infrequent flood events, with channels being dry most of the time. Irrigation works and similar human activities further complicate processes in such rivers.

This paper reports the results of an investigation of more than a century of channel changes in an example arid-region river, the Salt River of central Arizona, U.S.A. (Figures 1 and 2). Two specific research questions are addressed. First, how have channel locations, dimensions, and materials changed in response to floods? Second, how can observed changes be generalized?

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Figure 1. Channel of the Salt River near Tempe Butte looking west or downstream. A single main-flow or low-flow channel is outlined by vegetation in some areas. An artificially straightened reach occurs between the camera and the airport runway—it was destroyed and replaced by a slightly meandering channel in the flood of 1965. Photograph by Maddock Associates Aerial Surveys, held by the Phoenix Urban Study Office, U.S. Army Corps of Engineers (Frame 16, 15 September, 1962)

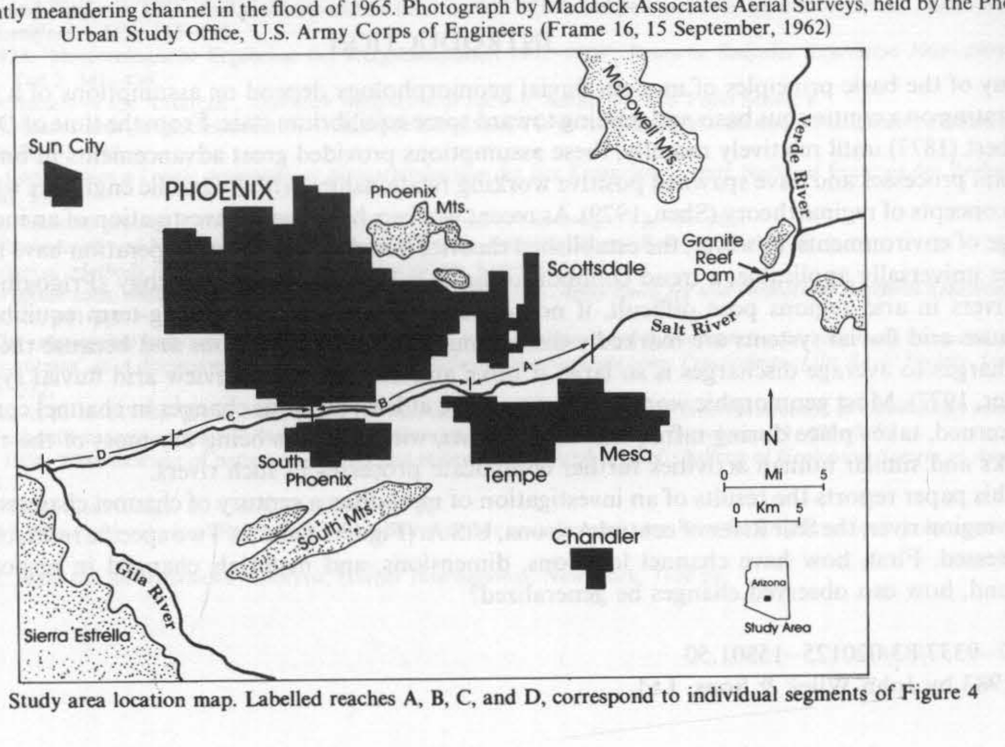

Figure 2. Study area location map. Labelled reaches A, B, C, and D, correspond to individual segments of Figure 4

The Salt River originates in the White Mountains of eastern Arizona from orographic precipitation generated by peaks up to 3533 m (11,590 ft) above sea level. The Salt River flows westward through mountain gorges until it emerges into the fault block valleys of the Basin and Range geomorphic province (see Hunt, 1974, for a general regional description; see Aldridge, 1970, for a description of the Salt River basin and its hydrology). At its confluence with the Gila River, the Salt River drains about 38,400 km^2 (15,000 mi^2). On the valley floors the channel only rarely encounters bedrock, and on the surface of alluvium several thousand metres thick it develops an unstable braided channel. The specific reach of the river analysed in this paper extends from a major irrigation diversion work, Granite Reef Dam, 48 km (30 mi) through agricultural and urban lands to its junction with the Gila River.

The channel might be characterized as braided, but it lacks the numerous subchannels of nearly equal magnitude found in some braided streams in glacial or semi-arid regions (Rust, 1972; Miall, 1977). The banks of the high-flow channel are poorly defined and are approximately 152 m (500 ft) to 1524 m (5000 ft) apart. Within these limits is a well-defined low-flow, invert, or main-flow channel. This main-flow channel has banks from 1 to 8 m (3 to 26 ft) high and a width ranging from 66 to 328 m (200 to 1000 ft). The main-flow channel is usually filled by flows that have a return interval under natural conditions of about 5 years. Channel materials range from coarse sand to very large cobbles and a few boulders with medium diameters of 0·6 m (2 ft) or greater. Although the channel has changed somewhat over the past century, it has not behaved like the nearby Gila River as described by Burkham (1972, 1976).

Limited portions of the mountain watersheds that feed the Salt River experience more than 62·5 cm (25 in) of precipitation annually (much of it as snow), but the majority of the basin receives only about 15 cm (7 in) (Sellers and Hill, 1974). Floods usually occur as products of intense winter rains or rapidly melting snow. The largest flood on record was 8400 $m^3 s^{-1}$ (300,000 $ft^3 s^{-1}$) (Table I). Substantial amounts of precipitation fall in the valleys from summer thunderstorms, but because of their limited size, these storms produce only local flooding. According to Powell (1893), the discharge of the Salt River before the imposition of upstream control structures was about 22·6 $m^3 s^{-1}$ (800 $ft^3 s^{-1}$), but dams now result in a dry channel except for flood periods (Central Arizona Water Control Study, 1980, discusses discharges).

Table I. Floods in the Salt and Gila River System in central Arizona

Year	Peak discharge m^3s^{-1} (ft^3s^{-1})
Feb. 1890	4004 (143,000)
Feb. 1891	8400 (300,000)
Apr. 1895	3200 (115,000)
Apr. 1905	3200 (115,000)
Nov. 1905	5600 (200,000)
Jan. 1916	3360 (120,000)
Jan. 1916	2940 (105,000)
Feb. 1920	3640 (130,000)
Feb. 1927	1960 (70,000)
Mar. 1938	2380 (85,000)
Mar. 1941	1120 (40,000)
Jan. 1966	1876 (67,000)
Feb. 1973	616 (22,000)
Mar. 1978	3416 (122,000)
Dec. 1978	3920 (140,000)
Jan. 1979	2464 (88,000)
Mar. 1979	1887 (67,400)
Feb. 1980	5040 (180,000)

Peak discharges on the Salt River at Phoenix from unpublished data provided by U.S. Geological Survey, U.S. Forest Service, and newspaper accounts. Indicative of relative flows in the study area. Unknown flow attenuation not accounted for.

LOCATIONAL CHANGES

The location of the centre line of the main-flow channel was determined for the entire study area for a variety of years from 1868 to 1980. Up to 13 different years of observation provided channel locational data for portions of the study area. Additional years of observation would be possible from the available historical data, but because preliminary investigations show that no significant changes took place before flood events, repetitive coverage between floods was unnecessary. Data on channel locations were derived from land surveys, topographic maps, irrigation maps, and aerial photographs. The centre line of the main-flow channel was plotted on base maps for subcomponents of the entire study area to generate a chronological picture of change.

A more meaningful perspective on channel locations is provided by locational probability maps, contoured maps showing the distribution of values that represent the probability of the channel being located at respective points. On such a map, for example, the 5 per cent contour line would connect all the points with a 5 per cent probability of being occupied by the channel. When applied to a portion of the Gila River downstream from the present study area, locational probability maps showed considerable spatial variation in probability values (Graf, 1981).

Construction of the probability map begins with the maps of centre lines for all the years of record superimposed on top of each other. Basic sample lines were constructed across the general flow area at 1·6 km (1 mi) intervals, and each sample line was divided into 150 m (500 ft) segments. The number of centre lines passing through each sample segment was tabulated and converted to a percentage of the total number of channels crossing the sample line. This percentage value was plotted at the centre of each segment of each basic sample line, and the entire distribution of values was contoured at a 10 per cent interval. Considerable smoothing results with the 1·6 km (1 mi) spacing, but a more detailed map could be obtained with more closely-spaced sample lines or by using a grid. Experimentation shows that different placement of the sample lines would result in different details emerging on probability maps, but the overall patterns would remain the same.

The locational probability map reduces a lengthy and complex record to a single, easily interpreted map with areas of stability and instability clearly defined. Stable sections of the general channel have high probabilities of channel location in a limited zone, indicating that the main-flow channel flows through the same sample segment most of the time. Unstable sections of the general channel have widespread low probabilities, indicating that the main-flow channel has passed through many different sample segments with few repeated patterns.

Locational probability maps have some disadvantages, however. They rely on historical data, and their accuracy is a function of the accuracy of that data which is frequently in a form not originally intended for this use. Length of record is critical, and for many rivers a century or more of data is not available. Because the maps are based on past positions of the channel, use of the maps for predictions of future locations assumes no significant changes in major controls such as climate, sediment supply, and human management. The final appearance of the maps is sensitive to the spacing of the original sampling lines, and smoothing of the data by contouring is somewhat subjective.

Changes in channel location along the 7·8 km (6 mi) reach of the Salt River from the Country Club crossing to the Mill Avenue crossing are representative of much of the study area (Figure 3). During the period from 1868 to 1926, wide fluctuations occurred in the lateral position of the channel, with lateral movements of about 1·5 km (0·9 mi) occurring near Country Club crossing. Between 1926 and 1961 much less movement was evident, but a special characteristic appeared in the 1937 channel: it was highly sinuous and exhibited an unusually pronounced meander near Tempe Butte. During this period phreatophyte growth was more dense than at any other time during the period of record. The density of the phreatophytes reduced channel capacity and induced sedimentation and channel avulsion (Hadley, 1961). From the mid-1960's to 1980 the channel migration was relatively minor despite the six major floods including the third largest of record. Most of the channel stability in the later part of the record is probably due to intensive degradation of the main-flow channel that began in 1965 and continued in subsequent floods. Gravel mines in the channel contributed to this down-cutting.

The locational probability maps for the main-flow channel provide a time-transgressive view of channel positions (Figure 4). In the reach from the Country Club crossing to Rural Road (Figure 4A) there is a large

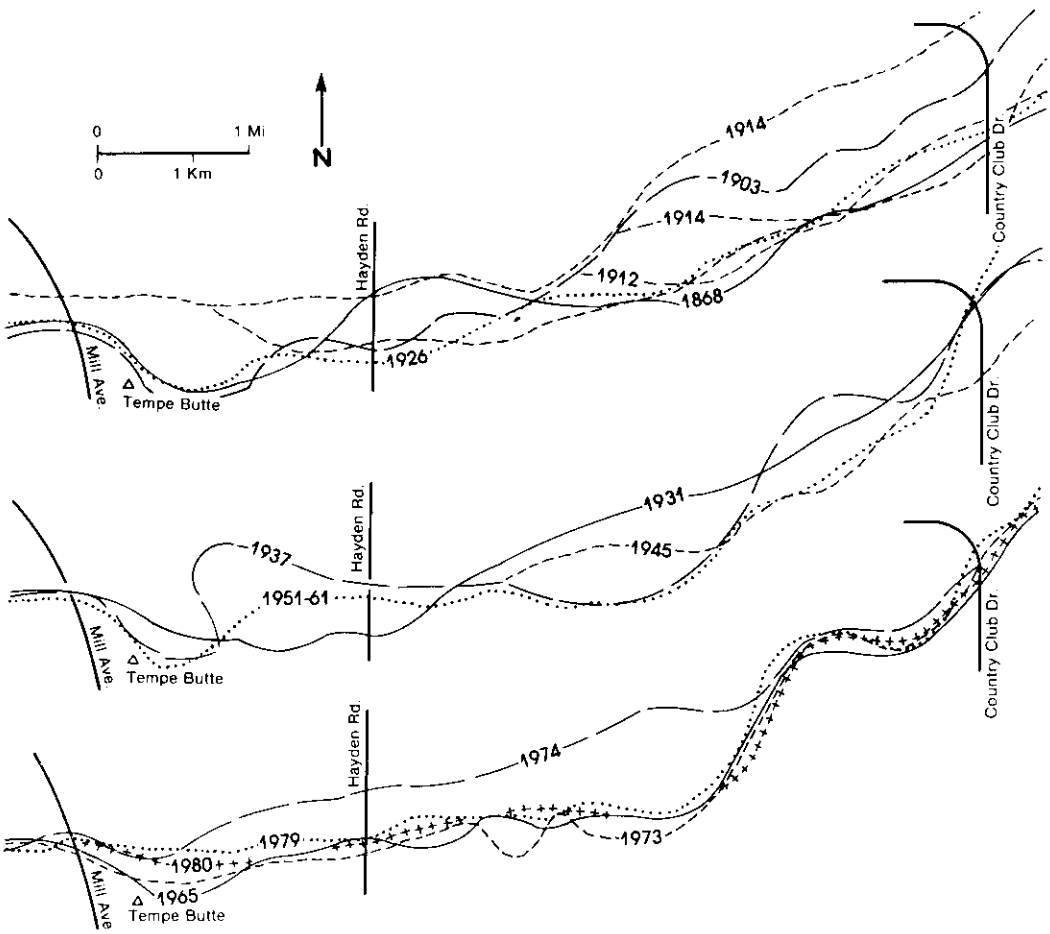

Figure 3. Historical locations of the main-flow channel in the Country Club crossing to Tempe Butte reach of the Salt River (reach A on Figure 2)

area of low probability, indicating instability. Near Tempe Butte a zone of relatively high probability begins, a product of the control exerted by the emergence of bedrock in the channel. Locally known as the Tempe Narrows, this bedrock portal is formed by remnants of weathered Precambrian quartzite and Tertiary andesite (Wilson *et al.*, 1957; Wilson, 1962). The zone extends about 2 km (1·2 mi) downstream (onto Figure 4B) before it begins to dissipate. By the time the river reaches the I-10 highway crossing, it has returned to the low locational probability configuration. Further downstream (Figures 4C and 4D) zones of stability and instability alternate with each other along the river with subregular spacing averaging about 3·2 km (2 mi).

Zones of high locational probability, and thus relative stability, appear to be associated with three features. One is the surfacing of bedrock in the channel floor at Tempe Butte. Other stable zones are co-located with engineering works, such as the stabilized location associated with the Central Avenue bridge (shown on the right side of Figure 4C). Finally, several zones of high locational probability appear to be products of an inherent sinuosity in the channel. If the channel maintains a relatively consistent sinuosity throughout the record, and if some channel locations are unchangeable because of bedrock or engineering controls, the channel has a limited number of options for location in those areas free of physical controls. Poorly defined meanders may swing to one side or the other, but the crossover points would almost always be occupied by the channel if sinuosity is to be maintained. An analysis of the main-flow channel over the 112-year record with 13 units of observation shows that the sinuosity for the total study reach varied from a maximum of 1·13 in 1937

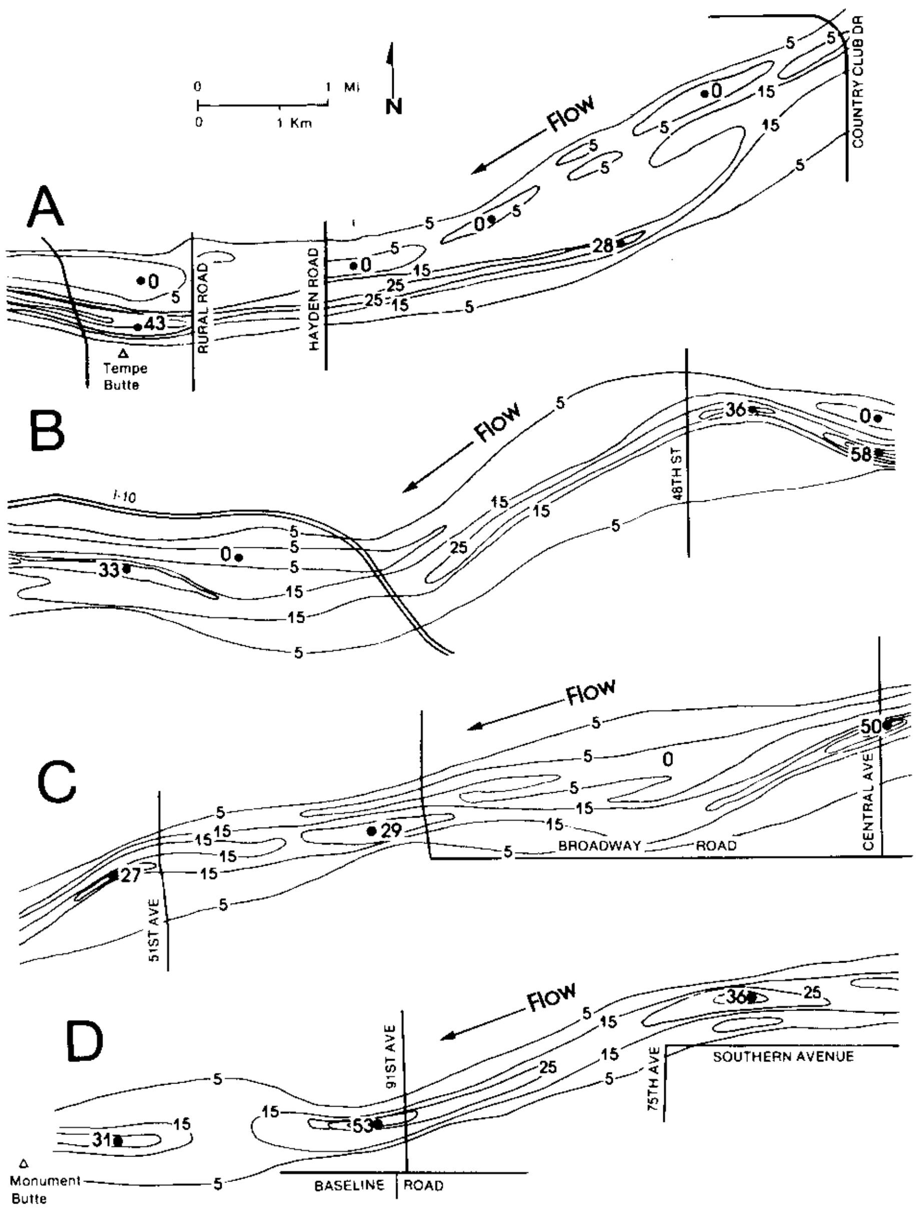

Figure 4. Locational probability maps for the main-flow channel of the Salt River. Segments are identified by letters on Figure 2 for location

to a minimum of 1·05 in 1868 and again in 1980 (sinuosity defined as the ratio of actual channel length divided by the shortest downvalley distance; Schumm, 1977). The mean is 1·08, with a standard deviation of 0·02 (Table II). Similar analysis for the nearby Gila River shows that there the mean sinuosity is 1·18. The difference between the two streams is the steeper gradient and larger particles in the Salt River. The consistency of sinuosity values through record is remarkable in view of the wide range in flood magnitudes (Table I) and the numerous human interferences with the process in the study area as well as upstream. In every case reaches of the channel that were artificially straightened re-established their sinuosity during the first flood following the straightening effort.

Table II. Channel sinuosity, 1868–1980

Statistic	Reaches * A	B	C	D	Combined total
Mean†	1·10	1·08	1·06	1·08	1·08
Standard deviation†	0·08	0·03	0·03	0·05	0·02

* Reaches keyed by letter to locations shown on Figure 2.
† Number of observations for each cell = 13.

CHANNEL DIMENSIONS

Sources of data for historical dimensions of the main-flow channel are less dependable than information on channel location. On historical maps cartographic conventions used to depict the channel margins are unclear: in some cases the main-flow channel appears to have been drawn, while in other cases the high-flow banks were defined or even high water lines from floods. Historical photographs (available in local library collections, State Historical Society in Tucson, National Archives and Library of Congress in Washington D.C., and the collection of the U.S. Geological Survey in Denver) for selected reaches provided insight into the implications of the maps. Between 1937 and 1980 aerial photographs provided reliable data on the width of the main-flow channel. Information on depth is also generally unreliable. Topographic maps were made at scales and with contour intervals that did not reveal the channel invert, and aerial photography is at inappropriate scales. Reliable data may be obtained through a few surveyed cross-sections (accomplished for construction purposes) and by means of historical photographs of bridges showing the piers in relationship to the bed.

Width measurements for 14 cross-sections of the main-flow channel for the period 1868–1980 show that the mean width was 125·3 m (411 ft) with a standard deviation of 26·8 m (88 ft). In the upstream portion of the study area widths are greater than the mean, commonly about 250 m (750 ft), while in the downstream area the width is much less than the mean. Though there were substantial fluctuations throughout the record in response to the 18 floods, there was no discernible trend toward either shrinkage or enlargement in the horizontal dimension. High-flow and over-bank flow areas have enlarged considerably in response to floods in the period 1978–1980, but the exact outlines of these features is highly dependent on near-channel construction activities entailed in urban development.

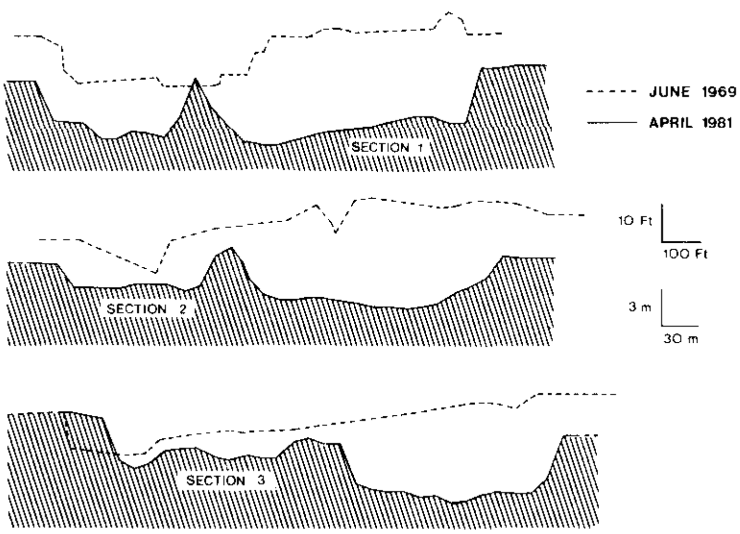

Figure 5. Channel changes at three sections approximately equally spaced between Hayden Road and Mill Avenue (see Figure 3). June 1969 configuration surveyed by Paul F. Ruff (unpublished, 1972); April 1981 and configuration surveyed by Herbert J. Verville (unpublished). During the period between the surveys the channel was dry except for the six most recent floods given in Table I. Shifts in main-flow channel location and degradation are evident

Between the time of the earliest available photographs of the river (the late 1880's) until the 1965 flood, the bed elevation of the Salt River in the study area changed very little. Numerous photographs show that the maximum vertical distance from bed to bank tops was less than 1·5 m (5 ft). Subsequent erosion has destroyed almost all of this bed, but a few scattered remnants, none larger than about 92 m^2 (1000 ft^2) in the surface area, confirm the photographic evidence in that all are about 1·5 m (5 ft) below bank tops.

Since the 1965 flood, much downcutting of the general channel and the main-flow invert has occurred (Figure 5). Most of the entrenchment occurred in the 1978 flood, with lesser amounts thereafter. The bed of the main-flow channel now lies approximately 6 m (20 ft) below the pre-1965 bed level. The gradient of the bank tops throughout the study area is 0·0019 (based on U.S. Army Corps of Engineers surveys), while the gradient of the pre-1965 channel was similar. The present channel gradient is 0·0017, an inconsequential change which suggests that the lowering of the bed has not been indicative of a gradient adjustment. An estimated $4{\cdot}5 \times 10^6$ m^3 ($1{\cdot}6 \times 10^8$ ft^3) of material has been excavated in the 48 km (30 mi) study reach during the seven floods from 1965 to 1980. This progressive entrenchment is probably responsible for the decline in channel mobility noted above for the later period record.

CHANNEL MATERIALS

The downcutting of the main-flow channel has proceeded through three distinct layers of sediment. The pre-1965 bed was in layers of course sand. Beneath the sand was a second layer of cobbles with medium diameters of up to 0·5 m (1·5 ft). During six floods erosion proceeded through the entire 4·6 m (15 ft) of the course unit, and the 1980 flood carried the erosion downward into the third layer, a mixed unit dominated by sands and gravels but with some larger particles as well. This continued downcutting indicates that the channel has the competence to transport the largest boulders in the bank/bed sediments in all areas of the channel disturbed, as well as in areas not included in the mining activity. Because the materials in the Pleistocene terraces along the river have particle sizes identical to those in the coarse unit of the bed (Kokalis, 1971), the present river must be similar to its Pleistocene predecessor in terms of its competence (Pewe, 1978).

The historical photographs and aerial photography mentioned above show a clear progression of differing bed materials through time which is directly related to the downcutting through various units. Figure 6 is representative of the changes. In 1949 (and extending back to the earliest photographs in the 1880's) the bed was predominantly sandy, with some cobbles probably transported into the study reach from mountainous areas upstream (Figure 6A). Although the diversion dam at the upstream end of the study area traps many sediments, it is silted in, and direct observation shows that small boulders are transported over the dam crest during floods. By early 1980 the bed was covered with cobbles that had the appearance of an armoured layer (Figure 6B), but this material was actually bedload as incision proceeded through the coarse unit. By the end of the 1980 flood, the main-flow channel had incised through the course unit and into the finer sediments below, so the bed was no longer a continuous sheet of cobbles (Figure 6C).

Sand and gravel mines in the channel provide an unusual opportunity to view the bedload transported during the 1980 flood. Because the channel was dry from 1941 to 1965, many mine operators established excavations directly in the main-flow channel. One pit (Figure 7) was completely filled by trapped bedload during the 1980 flood. It was re-excavated by mining operations, revealing the sequence of bedload deposits (Figure 8). The pit had been partially filled by sand during the early stages of the flood, when discharges up to 420 $m^3 s^{-1}$ (15,000 $ft^3 s^{-1}$) entrained sand left on the upstream bed during the waning period of the previous flood. When the discharge increased to 5040 $m^3 s^{-1}$ (180,000 $ft^3 s^{-1}$) large particles were brought into the pit, along with trunks of cottonwood trees (*populus fremontii*) that were entrained from the banks. As the flood continued, an increasingly wide range of particle sizes filled the pit to the top. At peak discharge when water depth was about 5 m (16·4 ft) above the bed level, particles were apparently in motion to a depth of about 7 m (22·8 ft), and as the peak waned the mixed sediments were deposited. As the flood stages declined, a sheet of sand was deposited on top of the sequence.

Evidence from the mine correlated well with evidence from the I-10 highway bridge, where a pier extending 8·2 m (27 ft) below the dry bed surface was undermined by particles in motion during the flood. Water depths and turbulence were slightly greater at the bridge site than in the mine area.

(A)

(B)

(C)

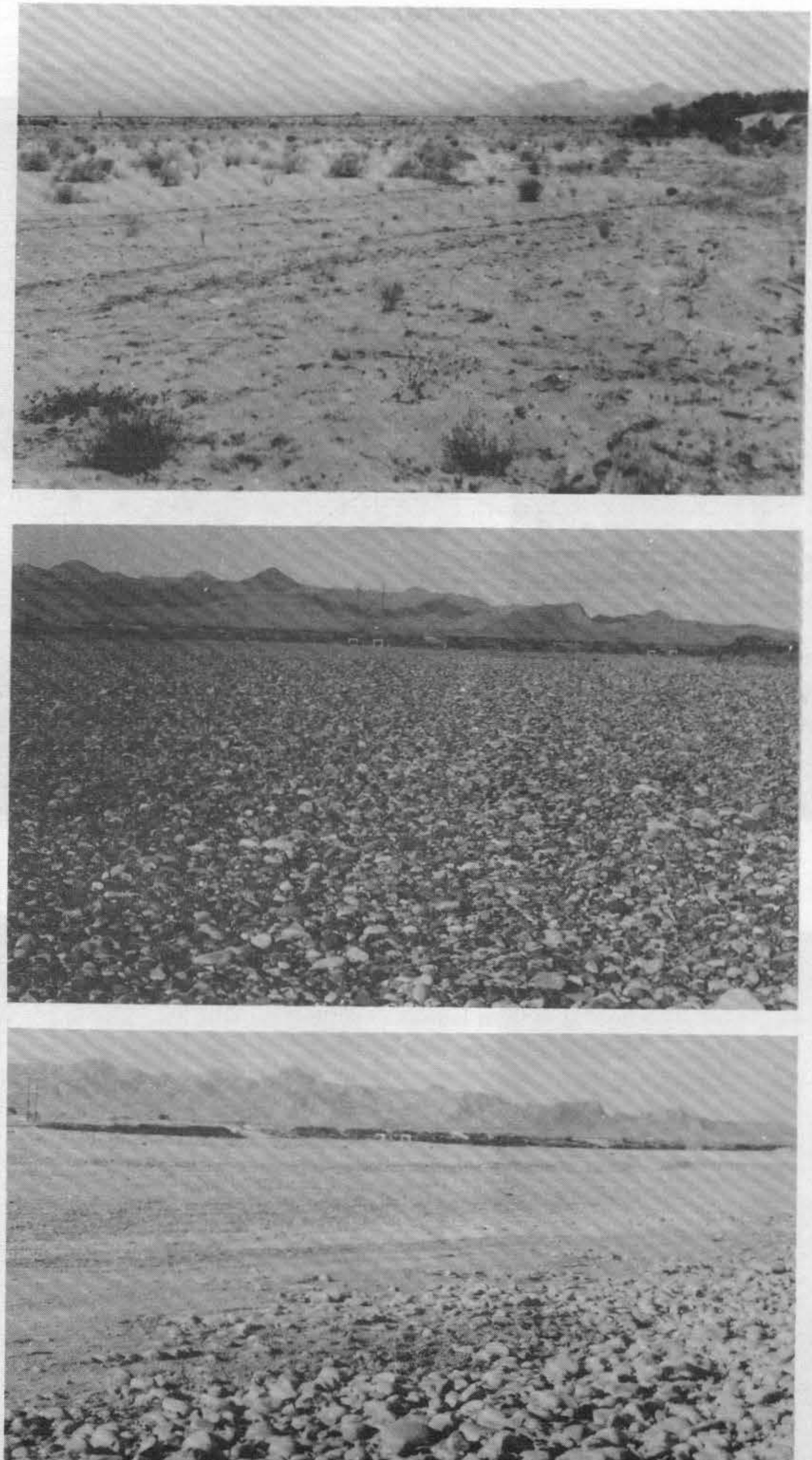

Figure 6. Sequential photographs showing changes in bed material; all looking upstream at a point about 12·8 km (8 mi) west of Granite Reef Dam with the McDowell Mountains in the background. View A: September 1949, showing a generally sandy bed (U.S. Bureau of Reclamation photograph, held by Phoenix Urban Study Office, U.S. Army Corps of Engineers). View B: January 1980, showing the cobble bed resulting from downcutting from five floods into coarse sediments beneath the 1949 bed. View C: December 1980, showing continued downcutting on the left resulting from the February 1980 flood into the layer beneath the coarse unit exposed in View B

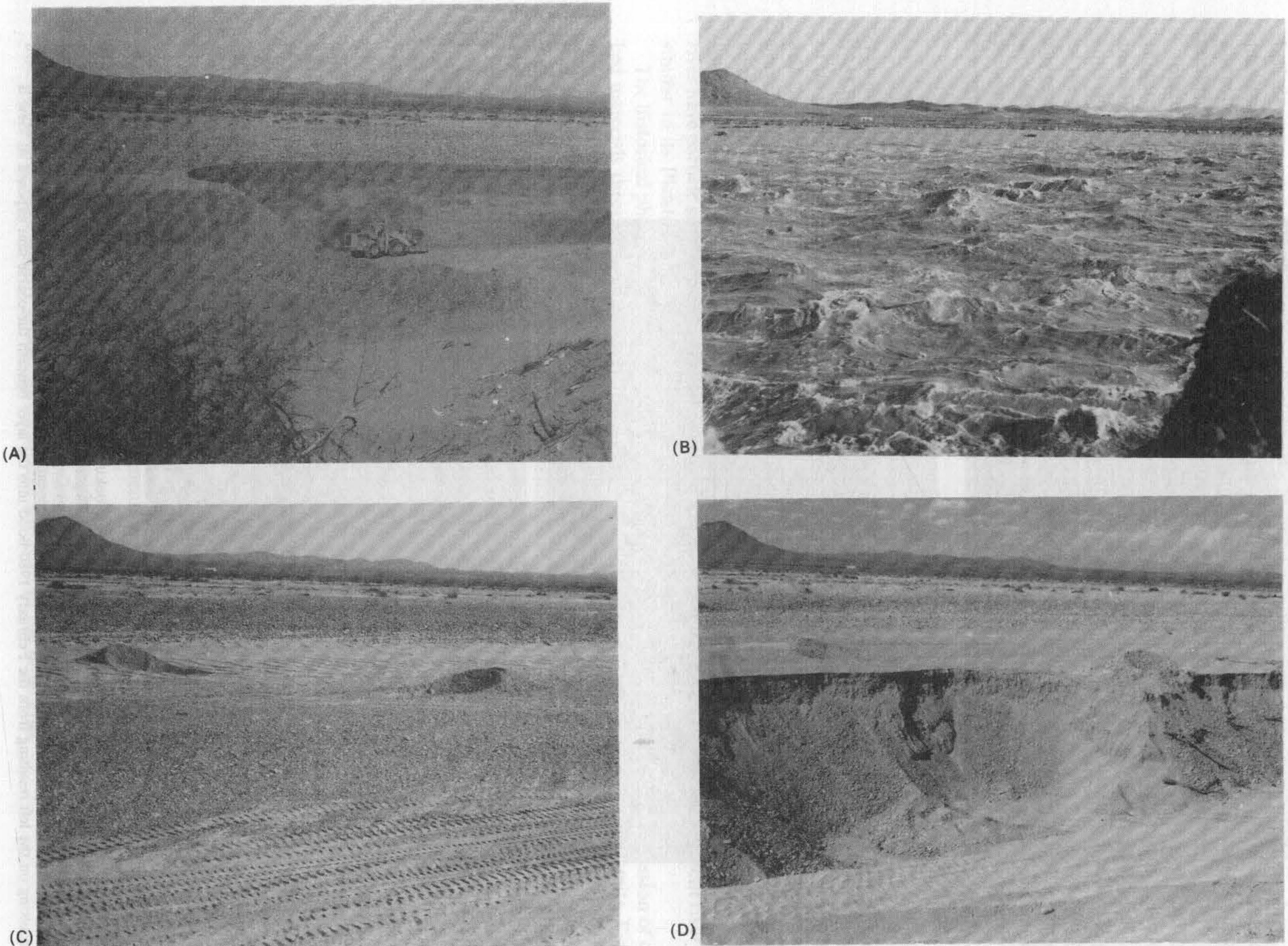

Figure 7. Sequential views of a gravel mine located in the main-flow channel of the Salt River about 8 km (5 mi) west of Granite Reef Dam. View A: January 1980, showing the mine excavated to a depth of 12 m (40 ft). View B: February 1980, showing turbulent flow of 5040 $m^3 s^{-1}$ (180,000 $ft^3 s^{-1}$) over the mine site. View C: April 1980, showing the complete refilling of the original mine by bedload deposition. Small excavations are sand and gravel prospects bulldozed by the mine owner. View D: October 1980, showing the re-excavation of the mine with a depth similar to the original. Cottonwood logs protrude from the sediments on the right

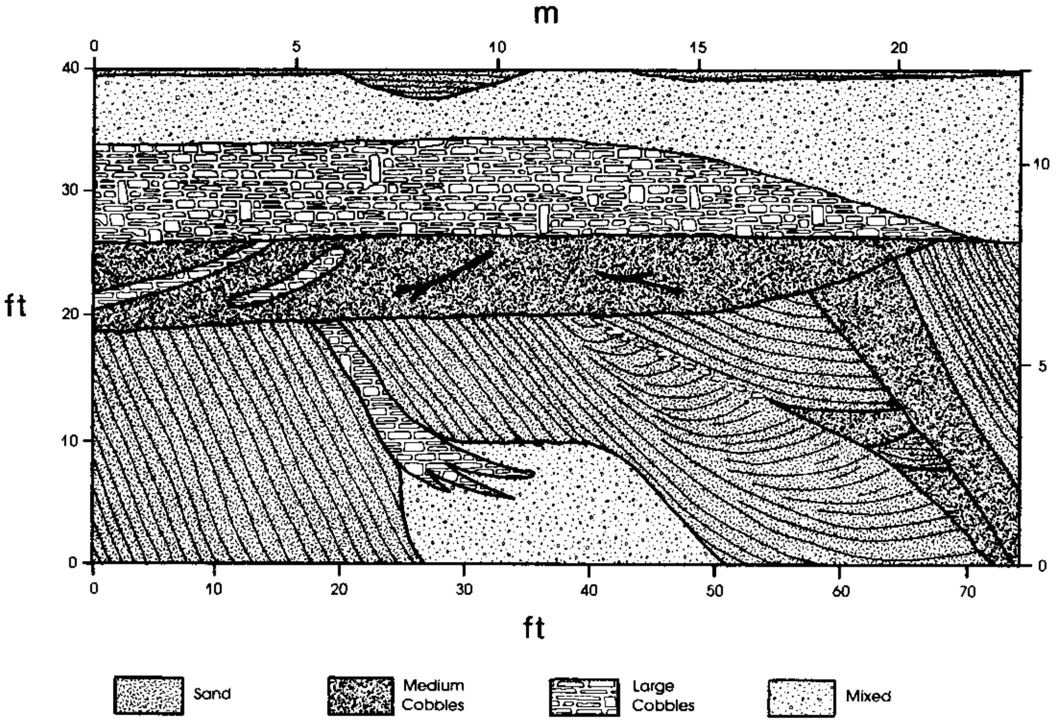

Figure 8. Diagram of deposits that filled a gravel mine during the February 1980 flood and later re-excavated. Same mine as shown in Figure 7. Flow from left to right

GEOMORPHIC ROLE OF FLOODS

Because all significant channel changes occur during floods, an understanding of channel dynamics relies on the definition of a threshold discharge of instability (see Schumm, 1979, for a discussion of thresholds). Discharges below this threshold value produce no channel migration or downcutting, while those above the threshold value result in channel destabilization. There are two approaches to definition of the threshold: deductive calculations and direct observation.

A flexible definition of the threshold of instability is afforded by calculations based on recognized relationships among discharge, channel dimensions, and the transport of coarse bedload. The determination proceeded as follows. Given the equation of continuity (Leopold *et al.*, 1964):

$$Q = WDV \tag{1}$$

where Q = discharge, W = channel width, D = channel depth, and V = velocity of flow. Also given the Manning Equation (in metric formulation with the coefficient of 1·009 simplified to 1·0; Derbyshire *et al.*, 1979):

$$V = D^{0\cdot67} s^{0\cdot5} n^{-1\cdot0} \tag{2}$$

where V = velocity of flow in m s^{-1}, D = depth of flow in m, substituted for hydraulic radius in the case of wide, shallow streams such as the Salt River, s = dimensionless gradient, and n = Manning's roughness coefficient. Substituting equation (2) into equation (1) for velocity and combining terms produced

$$Q = WD^{1\cdot67} s^{0\cdot5} n^{-1\cdot0} \tag{3}$$

and then solving for D:

$$D = Q^{0\cdot6} W^{-0\cdot6} n^{0\cdot6} s^{-0\cdot3} \tag{4}$$

Then given the Duboys Equation for tractive force or shear stress at the channel bed (Bogardi, 1974):

$$\tau_0 = \gamma D s \tag{5}$$

where τ_0 = tractive force in Newtons 0 kg m^{-2}, γ = unit weight of water or 1000 kg m^{-3}, D = depth of flow in m (substituted for hydraulic radius in the cases of wide shallow streams such as the Salt River), and s = dimensionless gradient. The term tractive force is used here because, as originally defined by P. Duboys in 1879, equation (5) described the 'force d'entrainement,' which is closer to its present application than 'shear stress'. The units for shear stress and tractive force are the same, however, and the choice of terms does not appear to be critical. Substituting equation (4) into equation (5) for depth, including the value for γ, and combining terms produced

$$\tau_0 = 1000\, Q^{0 \cdot 6}\, W^{-0 \cdot 6}\, n^{0 \cdot 6}\, s^{0 \cdot 7} \tag{6}$$

Baker and Ritter (1975) used a wide range of data to generate an empirical relationship between coarse bed materials and the amount of tractive force required to initiate motion and instability:

$$d = 65\, \tau^{0 \cdot 54} \tag{7}$$

where d = medium particle size in mm and τ_0 = tractive force in kg m^{-2}. Substituting equation (6) into equation (7) for tractive force and reducing to simplist terms produced

$$d = 2709 \cdot 85\, Q^{0 \cdot 32}\, W^{-0 \cdot 32}\, n^{0 \cdot 32}\, s^{0 \cdot 38} \tag{8}$$

Equation (8) can be used to determine the size of particles likely to be destabilized and entrained on the bed of a channel given discharge, width, roughness, and the slope of the channel. If, however, the known quantities include particle size, and the unknown quantity is the threshold discharge of instability, equation (8) may be solved for Q:

$$Q = (1 \cdot 87 \times 10^{-11})\, d^{3 \cdot 125}\, W n^{-1 \cdot 0}\, s^{-1 \cdot 1875} \tag{9}$$

The units are Q in m^3 s^{-1}, d in mm, W in m, n and s dimensionless. To convert the results of ft^3 s^{-1}, multiply by 35·71.

Solution of equation (9) produces a series of relationships between channel characteristics on one hand and threshold discharge of instability on the other (Figure 9). Solution of the model for the upper reach of the study area (Figure 4A) provided the following inputs: medium particle diameter was 130 mm, slope was 0·002, roughness was estimated as 0·050, and channel width was 229 m (750 ft). The calculated threshold of instability from these inputs is 545 m^3 s^{-1} (19,451 ft^3 s^{-1}). Based on the flood record and a Gumbel statistical distribution for event magnitudes (Ward, 1978), this discharge has approximately a five-year return interval.

The calculated value of the threshold discharge falls within the envelope of acceptable values defined by direct observation. During the 1973 flood with its peak discharge of 616 m^{3}s^{-1} (22,000 ft^{3}s^{-1}) significant channel migration occurred (shown on Figure 4A). During the early phases of the 1980 flood when the discharge was 420 m^3 s^{-1} (15,000 ft^3 s^{-1}) for several days, little channel adjustment occurred, and there were no relocations of the main-flow channel. Therefore, direct observation suggests that the threshold value must be between these two values, confirming the accuracy of the calculations in a broad sense.

Throughout this discussion the threshold for instability is equivalent to the threshold for initiating the transport of coarse bedload. This assumption is probably true for the Salt River and similar arid-region rivers, but it is probably not true for humid-region rivers and those dominated by suspended sediment transport.

CONCLUSIONS

Observations of the behaviour of the Salt River show that explanation based on concepts of equilibrium may be severely limited. Many aspects of the channel appear not to be responsive to varying inputs of water and sediment. Some parameters of the channel seem almost constant, irrespective of major changes in bed material. Once the discharge exceeds approximately 560 m^3 s^{-1} (20,000 ft^3 s^{-1}), all the bank and bed materials are

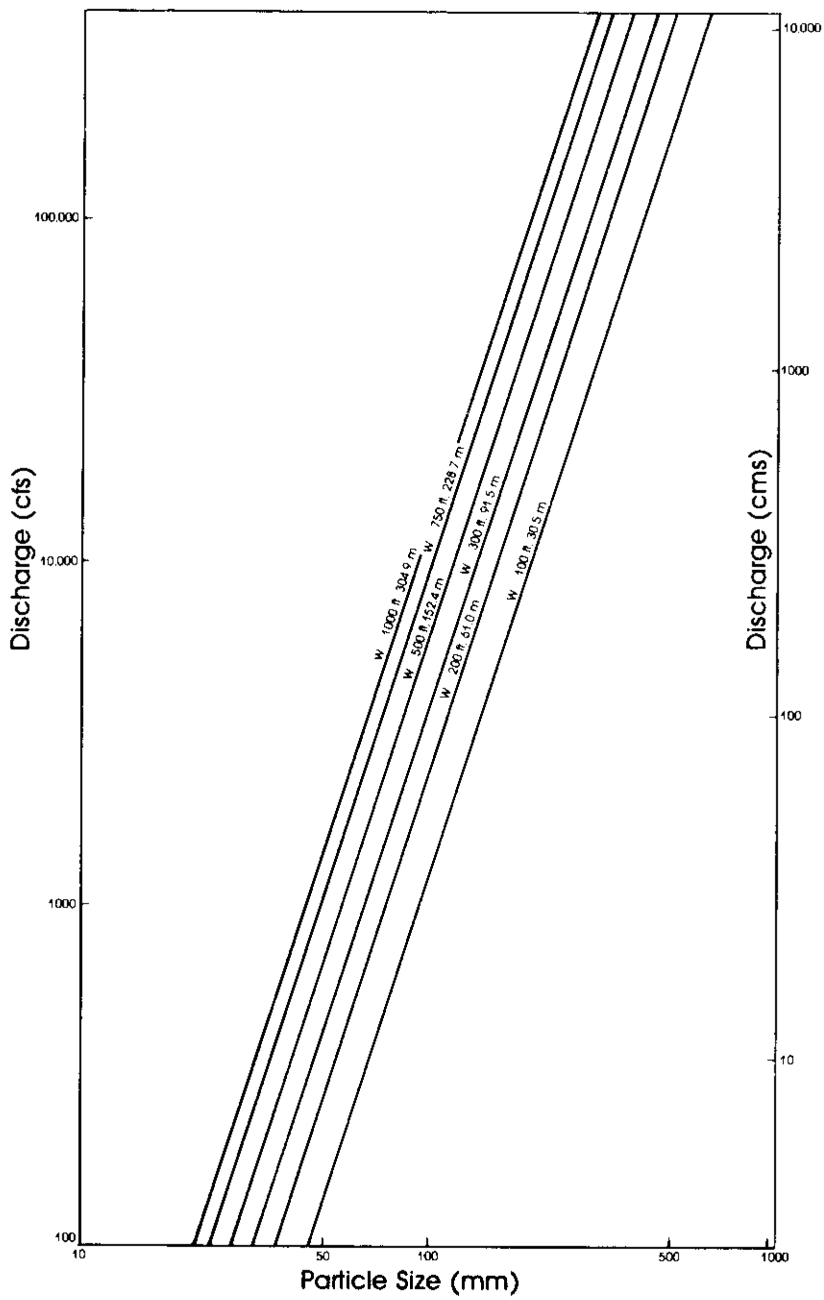

Figure 9. Threshold discharges of instability for various particle sizes and channel widths, calculated from equation (8) in the text calibrated for the Salt River in reach A as located on Figure 2

unstable. At discharges below that value, no adjustments take place. Channel width does not appear to be responsive to changes in discharge and sediment regimes, while depth of the main-flow channel seems more sensitive. Similar problems in application of equilibrium concepts and regime theory for the nearby Gila River have been noted previously (Stevens *et al.*, 1974), and in other areas by Graf (1979) and Thornes (1980).

A spatial perspective of the channel is best served by a probabilistic approach, with the channel space partitioned into probability zones. In the case outlined above, the probability values pertained to the likelihood of occupation by the active channel, but similar maps might depict probability of destabilization or some other spatial distribution related to process.

Data from 112 years of record for the Salt River provide tentative answers to the two research questions with which this paper began. First, floods have caused channel relocation and rearrangement with lateral migration up to 1·6 km (1 mi), but generally preserving a mean sinuosity value of 1·08. Recently, a series of increasingly large floods has caused deepening of the main-flow channel but relatively little widening. Channel bed materials have changed as excavation has proceeded through different layers of alluvium, but even boulders with medium diameters of 0·5 m (1·5 ft) have not formed armoured beds.

In answer to the second question, the observed changes can be generalized as locational probability maps which lead logically to spatial definition of alternating stable and unstable zones along the general flow area. Stable zones are either anchored by bedrock outcrops or engineering works or are defined by the combination of anchored points and a tendency for maintenance of a given sinuosity. Equilibrium concepts implying a balance among water, sediment, and channel dimensions are of limited utility, at least on a time scale of a century.

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