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PALEOHYDROLOGY OF THE EOCENE BALLENA GRAVELS, SAN DIEGO COUNTY, CALIFORNIA

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ABSTRACT

Steer, B.L. and Abbott, P.L., 1984. Paleohydrology of the Eocene Ballena Gravels, San Diego County, California. Sediment. Geol., 38: 181–216.

The Ballena Gravels are remnants of a river system that flowed westward across the ancestral Peninsular Ranges during medial and late Eocene time. The Gravels (actually conglomerate) are channelized fluvial deposits that built westward as alluvial fan (Poway Group), submarine canyon (Scripps Formation) and submarine fan (Jolla Vieja Formation) depositional systems. Because the integrated sedimentary system contains distinctive Poway rhyolite clasts of limited geographic and temporal extent the now separated component formations are recognizable on the San Diego coastal plain and on the Channel Islands. Paleogeographic reconstructions suggest a transport distance of about 315 km. Multiple techniques analysis suggests the channel gradient in the San Diego area was 12–18 m km^{-1}. Stream velocity, based on a competent particle size of 52 cm, ranges from 2.5 to 4 m s^{-1}. Eight equations based on slope and velocity generated estimates of channel depth at flood stage that vary from 2.5 to 4.5 m.

Paleodischarge can be estimated from regime-type engineering equations that are based on gradient, depth, and grain size. Paleodischarge also is calculated using a technique based on stream length and drainage area. Values generated by multiple techniques suggest 2.33-yr flood discharges of about 275 m^3 s^{-1}. Extrapolation based on a logarithmic curve indicates 100-yr flood discharges around 27,500 m^3 s^{-1}.

Channel-width estimates for 1 to 2.33 yr floods range from 25 to 75 m. Calculations of seasonally dominant rainfall, based on runoff and temperature, vary between 50–75 cm annually. Runoff is estimated from discharge values and paleotemperatures are based on caliche type, salt-fractured clasts, and the immature clay mineral suite.

The Nueces River of Texas shares some of the same characteristics of the Eocene Ballena river. The Nueces has a highly varied discharge due to seasonally intense rainfall similar to that interpreted for the Ballena river. Several rivers flowing to the Atlantic Ocean off the eastern side of the Andes Mountains in southern Argentina may also represent modern analogues.

INTRODUCTION

The Ballena Gravels (actually conglomerate) were named by Fairbanks (1893), who thought these gold-bearing gravels might be remnants of an ancient river that

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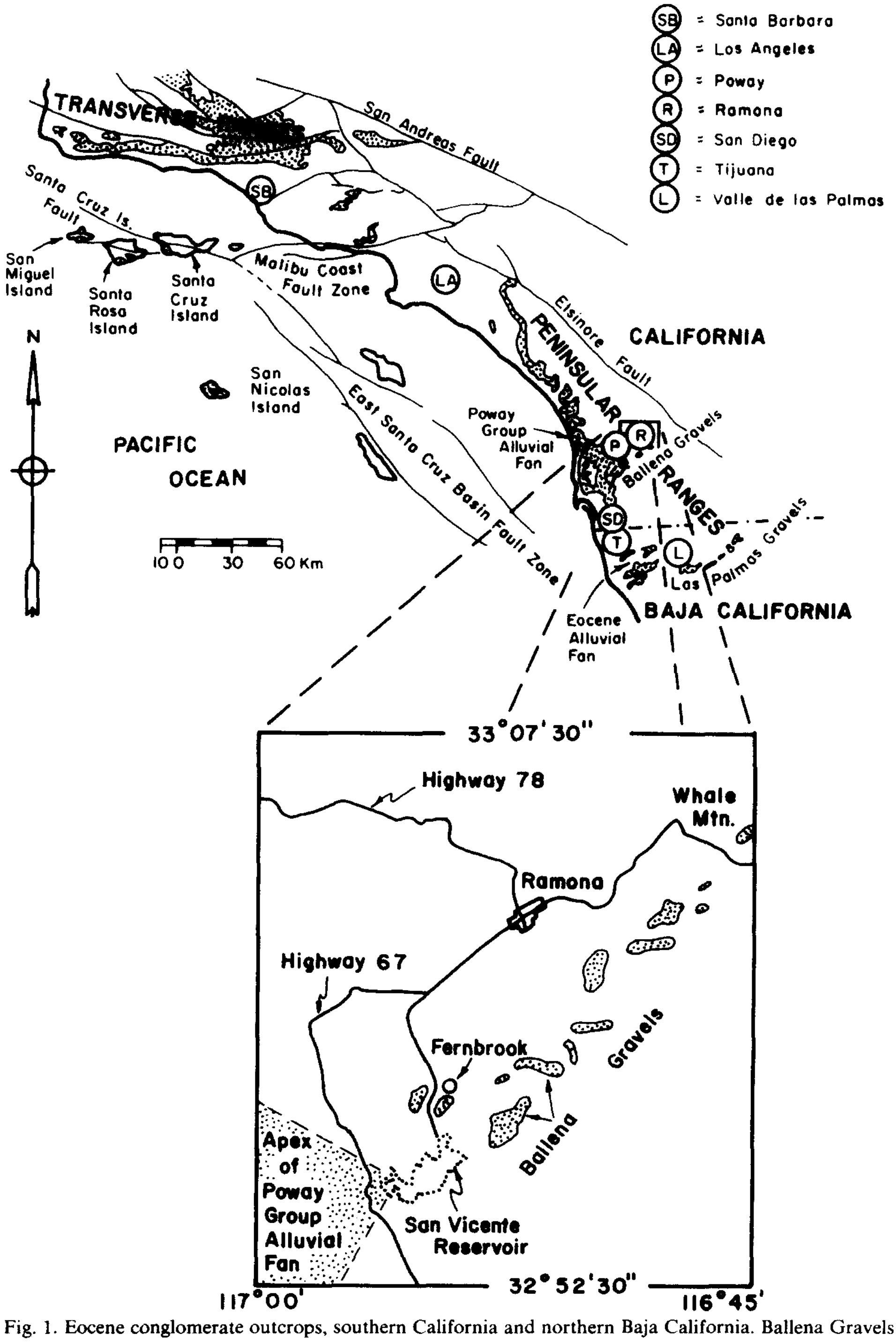

Fig. 1. Eocene conglomerate outcrops, southern California and northern Baja California. Ballena Gravels outcrops from Whale Mountain to San Vicente Reservoir.

flowed from an eastern source. Miller (1935), noting similarities between the Ballena Gravels and Eocene fluvial conglomerates in the Sierra Nevada, strongly favored a fluvial origin. He pointed out the similarities in clast assemblages of the Ballena Gravels and the Poway Conglomerate, an Eocene alluvial fan in San Diego, and suggested that the two rock units were deposited in the same system. These conglomerate clast suites in San Diego are dominated by distinctive, reddish-purple metarhyolites known as Poway clasts, that comprise over half of the gravels (Bellemin and Merriam, 1958; Woodford et al., 1968; Minch, 1972; Howell and Link, 1979; Kies and Abbott, 1983). The Poway rhyolite clasts are not found as bedrock in the Peninsular Ranges but were derived from outcrops well to the east.

The Ballena Gravels crop out discontinuously for 26 km in a southwesterly direction across the Peninsular Ranges (Fig. 1). They are remnants of an ancient river system that flowed westward during medial and late Eocene time through a valley cut into the widespread erosion surface that had nearly leveled the granitic core of the ancestral Peninsular Ranges. A S68°W flow direction subparallel to the outcrop trend is also shown by measurements of imbricated pebbles and cobbles (Minch, 1972).

The nonfossiliferous Ballena Gravels have been correlated with fossiliferous sedimentary strata of the Poway Group at San Diego on the basis of similar clast assemblages and nearly continuous outcrops. The easternmost exposure of Ballena Gravels is at Whale Mountain near the town of Ballena; westernmost exposures are found at San Vicente Reservoir near the Eocene slope break where gravels spread out westward as the alluvial-fan and fan-delta deposits of the Poway Group and the submarine canyon deposits of the Scripps Formation in the San Diego area (Fig. 1). The submarine fan deposits at the extreme western end of this system were offset later by strike-slip faults and now are exposed about 250 km to the northwest on Santa Cruz, Santa Rosa, and San Miguel Islands (Fig. 1).

The purpose of this study is to describe the paleohydrology of the Ballena river. Equations developed by engineers and fluvial geomorphologists are used to quantify ancient stream flow. Variables such as gradient, stream length, and elevation of source area were approximated from field characteristics and then were applied to the equations. This method has generated interesting estimates, however, the paleohydrologic regime presented is only an approximation.

FIELD CHARACTERISTICS OF THE BALLENA GRAVELS

Although the Ballena Gravels outcrops are only erosional remnants of the Eocene valley fill, they are up to 110 m thick and 3 km wide. In most outcrops, the Ballena Gravels unconformably overlie Cretaceous plutonic or older metamorphic basement rocks. Locally the Ballena Gravels overlie the Lusardi Formation, an extremely coarse boulder conglomerate of Late Cretaceous age (Nordstrom, 1970). The Lusardi Formation is composed of terrestrial fan and fluvial deposits. In some areas the

Lusardi stream channels were reoccupied by the Eocene Ballena channel (Minch, 1972). The Ballena Gravels are distinguished by the red-purple, Poway rhyolite clasts derived from afar that comprise over half of the gravel population. The exotic Poway clasts are totally absent in the granitic-dominated clast suite of the subjacent Lusardi Formation which was derived totally from local sources.

Locally in the Fernbrook and San Vicente Reservoir areas, the Ballena Gravels

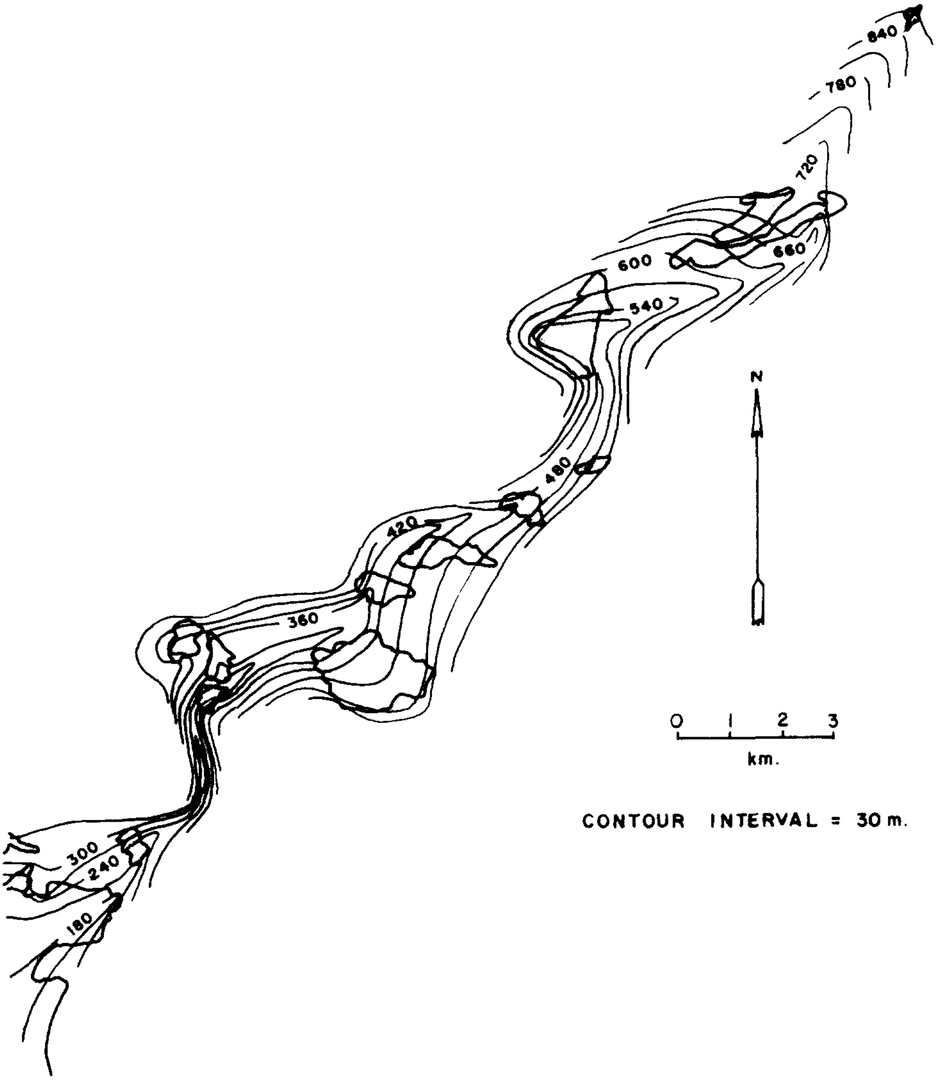

Fig. 2. Contour map on basal surface of Ballena river valley. Outlined areas are outcrops of Ballena Gravels. Topographic contours define Eocene valley eroded through Cretaceous batholithic rocks.

are underlain by green, sandy mudstone which, in turn, rests unconformably on basement rock, or the Lusardi Formation. The mudstone is up to 10 m thick and strongly resembles the middle Eocene Friars Formation of western San Diego County. Howell and Link (1979) interpreted these rocks as back-fill deposits formed during a transgressive event.

In plan view, the Ballena Gravels are strongly elongate with a sinuous pattern. The gravel beds are nearly horizontal and they abut granitic hillsides which are the Eocene valley walls. Minch (1972) contoured the Ballena channel by extrapolating between exposures and using the present topography as a guide. A contour map constructed using similar techniques is presented in Fig. 2. The pattern depicted by the contour lines approximates the extent of the paleovalley rather than a fluvial channel. The sinuosity of the paleovalley ranges from approximately 1.1 to 1.3. The overall slope of the basal surface of the Gravels at present is approximately 0.025 (25 m km^{-1}). This value is greater than the gradient of the ancient river because westward tilting of the Peninsular Ranges has increased the slope since the Eocene.

GRAIN SIZE ANALYSIS

Sieve analyses

Sampling sites had to be: (a) accessible by car due to the large sample size and heavy sampling apparatus; (b) sufficiently weathered to allow physical disaggregation of gravels and sands, but not weathered so severely that individual clasts crumbled; and (c) typical clast-size populations.

The minimum sample size was set at 135 kg. An area approximately 1 m^2 was randomly selected on a vertical outcrop and the sediment was disaggregated by hand or shovel. Data from the sieved samples were plotted as cumulative curves. Statistical parameters in Table I were derived from the formulas of Folk and Ward (1957).

TABLE I

Grain size parameters of the Ballena Gravels

	San Vicente Reservoir sample	Fernbrook sample
Grain size name	Cobble conglomerate	Cobble conglomerate
Median	-6.1ϕ (67 mm)	-5.5ϕ (45 mm)
Mode	-6.3ϕ (78 mm)	-5.8ϕ (56 mm)
Mean	-5.8ϕ (56 mm)	-5.2ϕ (37 mm)
Sorting	1.64ϕ	1.83ϕ
	Poorly sorted	Poorly sorted
Skewness	0.44	0.46
	Strongly fine skewed	Strongly fine skewed
Kurtosis	2.06	1.54
	Very leptokurtic	Very leptokurtic
Percentage of sand	9	12

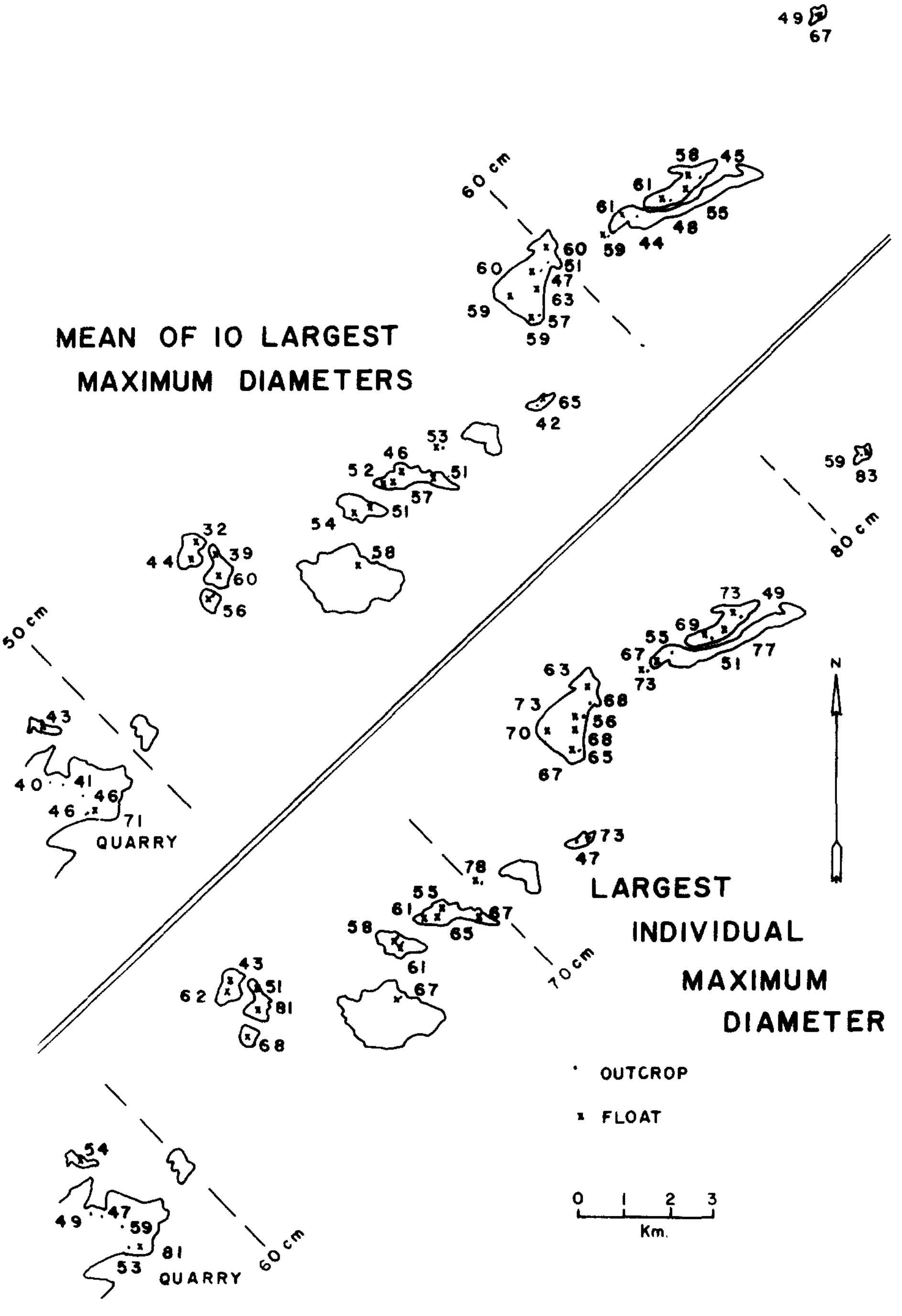

Fig. 3. Maximum diameters of exotic Poway rhyolite clasts plotted on base map of Ballena Gravels outcrops from Fig. 2. Upper plot is mean of ten largest maximum diameters. Lower plot is of largest individual maximum diameter at each outcrop locality. Contours of sizes used to calculate the coefficient of size reduction (a).

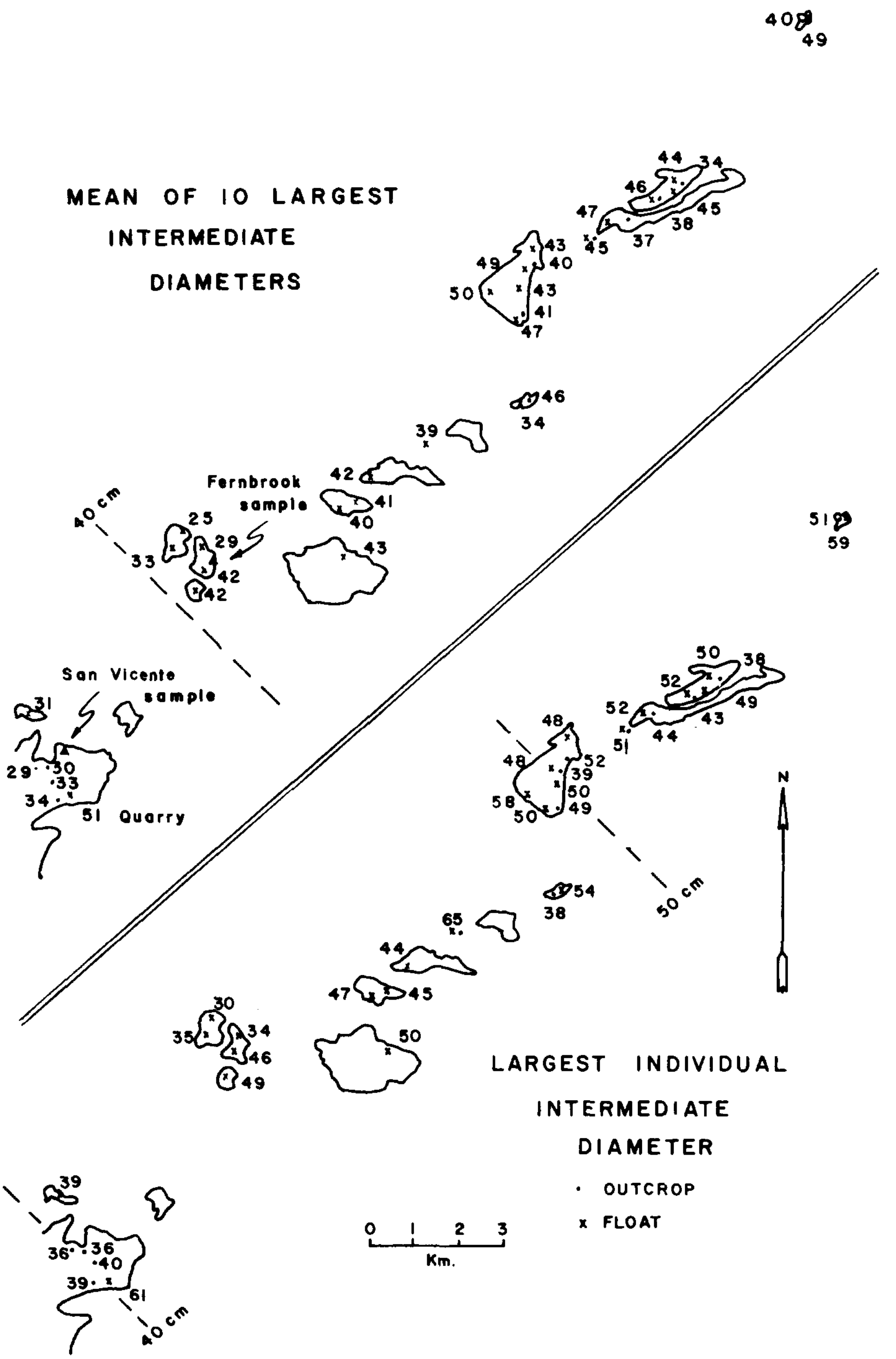

Fig. 4. Intermediate diameters of exotic Poway rhyolite clasts, Ballena Gravels outcrops. Upper plot is mean of ten largest intermediate diameters. Lower plot is of largest individual intermediate diameter at each outcrop locality. Contours of sizes used to calculate the coefficient of size reduction (a).

Maximum clast sizes

Grain-size data were obtained also by measuring both the intermediate and maximum diameters of the ten largest clasts found within a specific area. Both float (clasts in surface soil) and outcrop sample sites were used. Because the Ballena Gravels crop out mainly on isolated ridges or hill slopes where mixing from other rock units cannot occur, the float sites were regarded as valid for the evaluation of grain size. Only Poway rhyolite clasts were used in the study because their distant provenance (Abbott and Smith, 1978) insures that any size changes present reflect fluvial processes. Granitic clasts were contributed to the Ballena river throughout the outcrop area and thus their size changes are not necessarily related to downstream sorting or abrasion processes.

The mean maximum clast size (mean of ten largest clasts) offers a more reliable measure of sample size in coarse-grained deposits than mean grain size (Schlee, 1957; Meckel, 1967; Lindholm et al., 1979). Grain-size studies of this type have been conducted using both mean maximum diameter (Lindholm et al., 1979) and mean intermediate diameter (Bradley et al., 1972). In order to determine which length parameter is most strongly related to particle weight, random clasts from the Ballena Gravels were measured along the three principal axes and then weighed. A linear regression line generated from the data showed that intermediate particle diameter plotted against the natural log of particle weight had the strongest correlation (coefficient of 0.92). Maximum diameter also had a strong correlation (cf = 0.91). The minimum axis showed less correlation (cf = 0.87) and seems to reflect more strongly irregularities in the shape of the stones.

Measurements of intermediate and maximum diameters with generalized isopleths of grain size are plotted both as largest individual and mean of ten largest measurements in Figs. 3 and 4. Clast size decreases from northeast to southwest. The largest single clast (83 cm), as well as the largest value for mean maximum clast size (67 cm), are found at the easternmost exposures on Whale Mountain. Clast sizes taken from the westernmost Ballena Gravels are consistently smaller.

Much of the variation may be the result of comparing data collected at outcrop localities to data collected at float localities which are both plotted together on the figures. Measurements from outcrop localities are consistently smaller than those from float samples nearby. This trend was expected because weathering of the Ballena Gravels selectively enriches the float localities in coarser stones. Float sample values may be a better indicator of size variations in the Ballena Gravels because outcrop localities were commonly restricted to a relatively small number of stones. The generalized isopleth lines are based on the largest clast values and are not influenced by the smaller values.

PALEOHYDROLOGY

Introduction

Studies of both modern and ancient fluvial deposits have produced relationships that allow estimates of river parameters based on empirical and theoretical relationships among bed-material size, velocity, water depth, discharge, slope, and other flow parameters. We use estimates of channel gradient, stream length, sinuosity, elevation of source area, temperature, longitudinal stream profile, and drainage area to calculate values for stream velocity, channel depth, discharge, channel width, precipitation, mean annual discharge, and runoff.

Gradient

A downstream decrease in gradient gives most stream profiles a concave-upward appearance. Logarithmic and exponential curves can be used to approximate the longitudinal profile. Shulits (1941) showed that because the grain-size decrease in a river follows an exponential curve, the river profile dependent upon that load should also be represented by an exponential curve. Chorley (1969) pointed out that due to discontinuities from changing discharge, caliber of bedload, or channel characteristics, the equations usually provide only an approximate fit.

This study used an equation developed by Hack (1957) for streams in Virginia and Maryland to estimate the longitudinal profile of the Ballena channel. Hack's general equation for a stream profile is given by:

$$H = kL^{(z/r)+1}/[(z/r)+1] + C, \quad \text{when } z/r \neq -1 \tag{1}$$

where H is total fall from the drainage divide in ft, L is the stream length in miles, and r, z, C, and k are constants.

The constant r is a positive number originating from the empirical relationship between stream length (L) and drainage area (Ad) and is given by the expression:

$$L \propto Ad^r \tag{2}$$

Leopold et al. (1964) found that values of r range from 0.5 to 1.0 and that a value of 0.65 is applicable to many drainage basins in the United States.

The value z is a negative number determined from an empirical relation which expresses slope (S) as a function of mean annual discharge (Qm) and is given by:

$$S \propto Qm^z \tag{3}$$

In this relation, z describes the concavity of the stream's longitudinal profile. For most rivers, gradients are steeper in the upper reaches in order to maintain velocity and competence to move coarse material. In the lower reaches where sediment is finer and discharge is greater, shallower slopes are sufficient for material to be moved. Therefore, as grain size changes rapidly so does the gradient. Eicher (1969) stated that larger negative values of z indicate greater rates of change in gradient with discharge and hence profiles with greater concavity. Figure 5 shows several stream profiles based on differing values of z. The constant C moves the curve of the profile vertically on the graph but does not affect height differences (Eicher, 1969). A value of zero is used for C in this study.

The factor k relates channel slope to stream length (L) in the following expression:

$$S = kL^{z/r} \tag{4}$$

The value of k can be determined by solving eq. 1 for k:

$$k = H[(z/r)+1]/L^{[(z/r)+1]} \tag{5}$$

This value of k is then substituted into eq. 4, which is used to estimate gradients at places along the course of the channel. In this study, eqs. 4 and 5 will be used to estimate the channel gradient of the ancient Ballena river in the San Diego area. Estimates of channel length, total fall, and values for z and r will be necessary to solve the equations for gradient.

Based on a Sonoran source for the Poway clast suite, an estimate can be obtained for the Ballena river length. Palinspastic reconstruction of the southwestern United States and northwestern Mexico, with the Gulf of California and Sonoran structural dilation removed, yields a distance of 200–300 km from the Sonoran source to the Ballena Gravels outcrops (Minch, 1972). A central value for the distance is approximately 250 km. This figure represents a straight-line distance from the Sonoran source to the distal Ballena Gravels depositional site. The actual channel length was

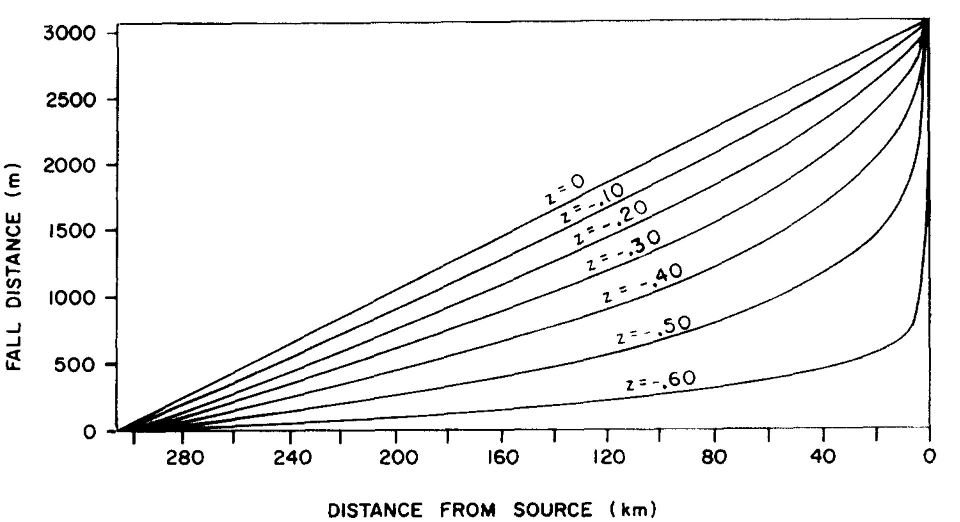

Fig. 5. Representative stream profiles for different values of z based on a total fall distance of 3000 m and a stream length of 315 km (after Eicher, 1969).

probably longer because of meandering intrinsic to fluvial systems. Contouring of the basal surface of the Ballena Gravels yielded valley sinuosities of 1.1 to 1.3. When these values are applied to the distance from the Sonoran source to the depositional site, they suggest a length for the Ballena channel of approximately 315 km.

The total fall distance also is necessary to calculate gradient from eqs. 4 and 5. The present elevation for the Sonoran source area ranges from 2000 and 3000 m for mountainous areas and from 1000 to 2000 m for valleys and alluviated areas. Minch (1972) pointed out that these values might represent minimum elevations for the Eocene source area and that the actual source-area elevations may have been much higher. If it is assumed that the Eocene arc of the southwestern U.S. and northwestern Mexico was broader and lower than the modern Andean arc, then a reasonable estimation for the average elevation of the arc might be somewhere in the 3000 m range. Based on these interpretations, and Minch's (1972) elevation estimates, a range from 2150 to 4000 m was used in this study as approximate average elevations of valleys and peaks in the Eocene Sonoran source area. For the Ballena river, the total fall distance (H) would have been effectively equal to the elevation of the source area because the depositional site of the distal Ballena Gravels was near sea level.

Leopold et al. (1964) showed that streams with large changes in grain size will have a tendency toward large negative values of z, as great as -1.65, whereas streams that have no significant decreases in grain size will tend to have z values closer to zero. As shown earlier, the Ballena Gravels have a gradual downstream decrease in grain size. Because size changes strongly influence concavity, and hence z, the Ballena Gravels would probably have a low negative z value, somewhere between -0.20 and -0.40 corresponding to a flattened longitudinal stream profile.

Leopold et al. (1964) have shown that the value r, which is related to stream discharge and stream length, varies from 0.5 to 1.0 and is controlled by climate. For humid regions, the exponent is closer to 1, whereas more arid climates would have r values closer to 0.5. It has been suggested that a semi-arid climate prevailed in the San Diego area during the time of deposition of the Ballena Gravels (Peterson and Abbott, 1979). Based on this assumption, an r value of 0.65 is used in this study.

Results of gradient calculations. When the estimated values of stream length, fall distance, and profile concavity are inserted into eqs. 4 and 5, a range of values can be calculated for the probable gradient of the Ballena river in the San Diego area. Figure 6 is derived from eqs. 4 and 5 and relates probable elevations in the Sonoran source area to the Ballena river slope at a distance of 315 km from the source. Different values for profile concavity (z-values) are represented by lines of different slope. On Fig. 6 probable and most probable areas are delineated based on estimations of fall distance and z-values.

The highest gradient value falling within the "probable" area is 21 m km^{-1} and represents a z-value of -0.20 and a fall distance of 4000 m. The lowest gradient, 7 m km^{-1}, is produced when the z-value is -0.40 and the fall distance is 2150 m. The "most probable" area is bounded by z-values of -0.20 and -0.40 and elevations between 2800–3700 m. Using these values, gradients from 12–18 m km^{-1} are calculated.

Velocity

An understanding of the velocities which existed in the Ballena river can help to define discharge as well as provide an indirect means of determining flow depths. Velocity is defined as discharge per unit area and is given by the following equations:

$$V = Q/A \tag{6}$$

and:

$$V = Q/WD \tag{7}$$

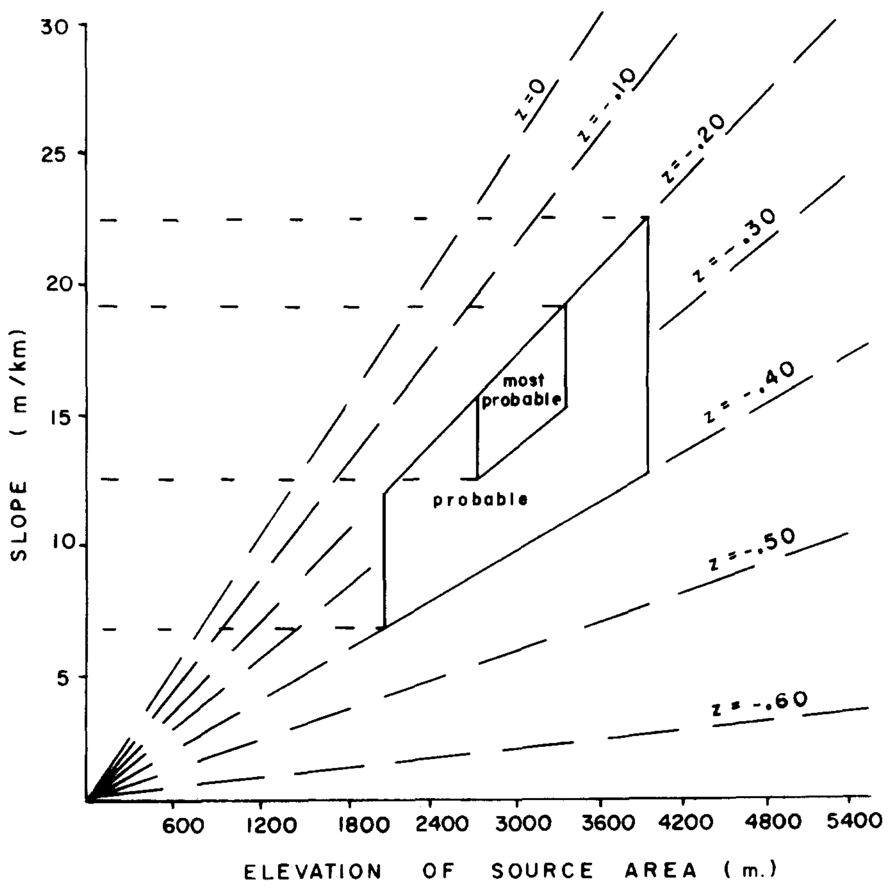

Fig. 6. Slope of the Ballena channel in the San Diego area based on 315 km long stream and estimated values of z and total fall distances.

where V is the average velocity, Q is the discharge, A is the cross-sectional area, W is the channel width, and D is the average depth. These relations have been used along with channel geometries measured in the field to estimate ancient flow velocities (Padgett and Ehrlich, 1976).

Sedimentary structures, such as cross-bedding and ripple marks, have been used to predict the velocities necessary for the formation of fluvial bedforms (Moody-Stuart, 1966; Hand et al., 1969; Cotter, 1971; Leeder, 1973; Nummendal, 1973; Koster, 1978). In the Ballena Gravels, channels and sedimentary structures are difficult to recognize because of limited exposures and gravel-on-gravel contacts produced by amalgamation.

Grain-size distributions can provide information on ancient flow conditions, especially velocity. The most common method used to estimate paleo-velocities is to determine the flow conditions necessary to initiate motion of the bed load (Malde, 1968; Baker, 1973b, 1974; Bradley and Mears, 1980).

Mavis and Laushey (1949) stated that velocity, or some function of velocity, near the channel bed is the main cause of the setting in motion of individual grains. Velocity is a function of the stream's competence which is defined as the largest particle a stream can move in traction as bed load. Competence is a measure of the stream's transporting ability and can vary along the course of the river as well as from time to time in the same area. In the gravel-dominated Ballena river, coarse sediment was abundant and thus competence can probably be estimated fairly accurately.

Methods. A particle-size value of 52 cm was used in eqs. 9–13 to estimate velocities of the Ballena river. This size value corresponds to the average intermediate diameter of the coarsest 10% of the clasts measured in the grain-size analyses. The largest clast measured had an intermediate diameter of 65 cm; however, clasts this size are rare and the lesser value may reflect stream competence more accurately.

Two basic methods have been used by researchers to study the conditions necessary for incipient sediment movement on a stream bed. They are expressed in terms of: (1) grain size and critical velocity; and (2) grain size and critical tractive force (shear stress). Many of these methods were determined by engineers and used in the design and construction of canals and waterways. Empirical and theoretical approaches have been used to study incipient grain motion. The empirical relations are derived from data obtained from studies of flumes, canals, and natural rivers. These data are commonly plotted on scatter diagrams that relate either velocity or shear stress to grain size. If the plotted points show a consistent relation, a regression line is generated to relate the points. The theoretical methods generally equate the forces of movement to those of resistance.

For this study, eleven different techniques were used to estimate paleo-velocity, seven relate grain size to velocity and four relate grain size to shear stress. A multiple methods analysis was employed wherein the results from techniques were compared to establish consistent answers. A similar procedure was followed by Bradley and Mears (1980) in determining paleo-velocities of Boulder Creek, Colorado.

Some of the equations yield velocities in terms of a critical bottom velocity which is directly responsible for the initiation of sediment motion. However, other equations give velocity as a mean flow velocity that depends on channel shape and depth and can be different from competent bottom velocities. In this study, competent bottom velocity (V_b) and mean flow velocity ($\bar{V}$) for average stream conditions are related by an equation given by Baker (1973b):

$$V_b = 0.7\bar{V} \tag{8}$$

Mavis and Laushey (1949) compiled data from earlier works and conducted flume experiments using sand and gravel to derive an empirical relationship between competent bottom velocity, particle size, and particle density. A regression analysis yielded the following equation:

$$V_b = \tfrac{1}{2} d^{4/9} (s-1)^{1/2} \tag{9}$$

in which V_b is competent bottom velocity in ft s^{-1}, d is diameter of grains in mm, and s is specific gravity of the sediment. Values used for specific gravity were 2.52 for rhyolites and 2.60 for quartzites (Telford et al., 1978).

Koster (1978) used a regression equation to estimate velocities from grain size on Spring Creek, Alaska. The equation is:

$$V = 49\, d^{0.381} \tag{10}$$

where V is mean flow velocity in cm s^{-1} and d is grain diameter in cm. The equation has a correlation coefficient of 0.825.

Miller et al. (1977) conducted flume experiments using quartz density material in water. One of their equations is:

$$V = 160.0 D^{0.45} \quad \text{for } d > 0.2 \text{ cm} \tag{11}$$

in which V is the velocity 100 cm above the bed (assumed to be approximately equal to V_b) in cm s^{-1} and D is the grain diameter in cm. This equation has a correlation coefficient of 0.99.

Bradley and Mears (1980) used two engineering equations derived by Torpen (1956) that were originally used for dam construction in Oregon. The equations are:

$$V_b = 4.1 d^{1/2} \tag{12}$$

and:

$$\bar{V} = 3.5 d^{1/2} \tag{13}$$

in which d is grain size in inches, V_b is bottom velocity for sliding material, and $\bar{V}$ is mean velocity for overturning rock. Both velocities are in ft s^{-1}.

Baker (1973b) pointed out that many of the formulae derived from engineering practices have specialized applications and difficulties may arise when attempting to approximate from geological evidence the variables in such expressions. Baker and Ritter (1975) stated that the main problem of using critical velocities is the extreme variability of the velocity profiles. Helley (1969) demonstrated that both particle shape and specific gravity have a direct influence on the initiation of particle movement. Variables such as these are difficult to isolate and are often in relations involving velocity and particle size.

Some workers have felt that critical tractive force (shear stress) is a more reliable measure of stream competency than those obtained from velocity relations. Several researchers have applied critical tractive force relations to problems of ancient flow conditions (Birkeland, 1968; Baker, 1973b, 1974; Baker and Ritter, 1975; Koster, 1978; Bradley and Mears, 1980).

A multiple techniques approach is used again to evaluate the critical tractive force relations. In order to compare the results, values of critical tractive force were converted to average velocities by an equation presented by Novak (1973):

$$V = 0.4\tau^{0.333} \tag{14}$$

where V is the average velocity in m s^{-1}, and τ is the critical tractive force in kg m^{-2}. Koster (1978) pointed out that a strong correlation exists between critical tractive force and mean flow velocity in this equation.

Baker and Ritter (1975) stated that a more suitable approach for geologic studies is to examine the relationship of intermediate particle diameter to critical shear stress. Baker and Ritter's compilation of data is shown in Fig. 7 and is equated by a regression line with the equation:

$$D = 65\tau^{0.54} \tag{15}$$

where D is particle size in mm and τ is the shear stress in kg m^{-2}. The regression line has a correlation coefficient of 0.92 and closely fits the points. To overcome the variance in the diagram, two lines were drawn encompassing 75% of the data points.

Abundant shear-stress data have been compiled by Fahnestock (1963), Church (1972), and Baker (1973b). Grain sizes are converted to a range of tractive-force values which are in turn converted to a range of average velocities by eq. 14.

Results of velocity calculations. Mean velocity values determined from eleven techniques are presented in Table II. The seven techniques relating velocity to grain size give a range of values from bottom velocity to average velocity. A range is given because water depths are unknown. If the stream was very shallow, the average stream velocities would be similar to bottom velocities. If the stream was much deeper than the dimensions of the bed material, average velocities might be about 1.4 times greater than bottom velocities.

Five of the techniques (Novak, Helley, Baker, Church, Fahnestock) also give velocity as a range which results from enclosing 75% of the data points within two lines parallel to a central regression line equating the points. The equation of Miller et al. (1977) yields the highest values of velocity (9.5 m s^{-1}). Their relationships are

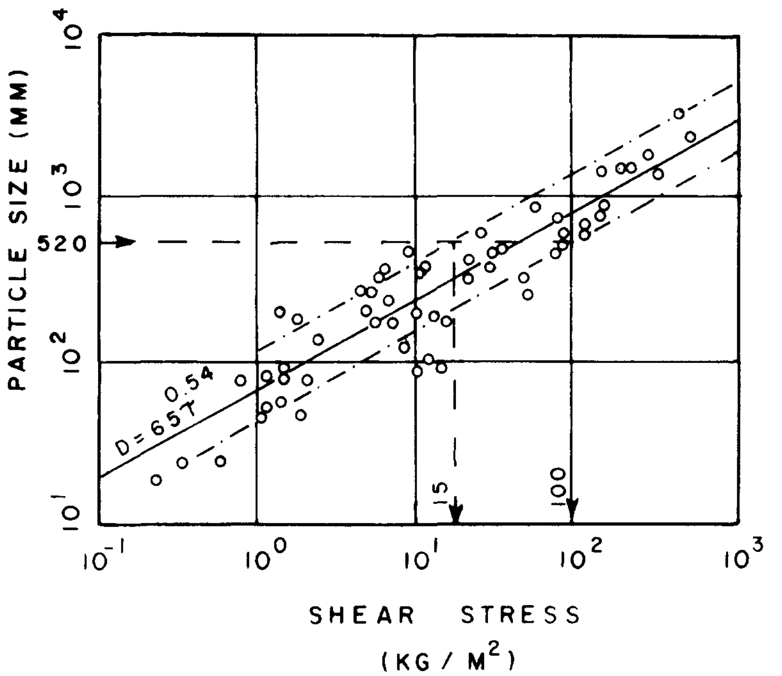

Fig. 7. Particle size versus mean shear stress for coarse bedload material. Shown are regression line, Shields criterion, lines encompassing 75% of data points, and shear stress for a 520 mm boulder from the Ballena Gravels.

TABLE II

Velocity results

	Shear stress (kg dm^{-2})	Mean velocity (m s^{-1})
Velocity – grain size techniques:		
Mavis and Laushey (1949)		3.1 to 4.3
Koster (1978)		2.2 to 3.1
Miller et al. (1977)		$V_{100} = 9.5$
Novak (1973)		2.0 to 4.0
Peterka et al. (1956)		3.5 to 5.1
Torpen (1956)		4.8 to 8.0
Helley (1969)		1.5 to 4.4
Shear stress – grain size techniques:		
Baker (1973b)	8.2 to 185	1.2 to 6.1
Shields (1936)	42.1	1.3 to 3.1
Fahnestock (1963)	12.4 to 41.2	1.1 to 3.4
Church (1972)	12.8 to 193	1.5 to 6.1

based on flume experiments that used sand-size particles with extrapolations into the cobble and boulder range. The extreme values obtained suggest that the results from this technique may be invalid. Engineering techniques by Torpen (1956) and Peterka et al. (1956) produce relatively high velocity values (8 and 5 m s^{-1}). They are also extrapolated from their original conditions but still give reasonable results.

Shear stress equations produced the lowest velocity values. Tractive force techniques by Baker (1973b) and Church (1972) gave the broadest range of velocity values after conversion. The broad range is related to the scattering of points which are presented as indiscriminate plots of unequal data. This type of plot has been criticized by several authors (Mackin, 1963; Baker and Ritter, 1975), but is still the most common method for displaying empirical stream data.

Table II shows a broad range of overlap within the velocity values. The greatest degree of overlap occurs at 3 m s^{-1} where eight of the eleven techniques are in agreement.

By using multiple techniques analysis to determine overlaps in the results, extreme values such as those determined from the techniques of Miller et al. (1977) can be disregarded. The range of velocity values predicted is significant because a high degree of agreement is obtained from a variety of different techniques that are based on both empirical and theoretical investigations. From the above data, it can be concluded that annual to 2.33 yr flood-water velocities in the ancient Ballena river probably ranged from 2.5–4 m s^{-1}. These velocities were sufficient to transport boulders with an intermediate diameter up to 52 cm.

Depth

Numerous techniques for estimating paleodepth are found in the literature. Some are based on direct measurements of paleochannels or sedimentary structures found in the field, whereas others have used stream characteristics such as velocity, channel gradient, and grain size distributions to make quantitative estimates of paleodepth. Since diagnostic channel features are generally absent in the Ballena Gravels, a quantitative approach using depth-related equations derived by engineers and fluvial geomorphologists is used herein to estimate paleodepth.

A multiple-techniques approach similar to that used in the velocity determinations has been applied to six different depth equations that are based on the parameters of velocity, gradient, bed shear stress, bed-material size, and fluid and particle density. Sandstone lenses found in the Ballena Gravels are also used to estimate paleodepth.

Velocity values used in the computations ranged from 2.5–4 m s^{-1}, and gradient values ranged from 7–21 m km^{-1} (0.0025–0.0083). The average particle size taken from cumulative curves is 6 cm and the particle size corresponding to the coarsest 10 percentile is 25 cm.

The Manning equation has been used to estimate flow velocities, channel gradients, and channel depths in ancient river systems (Cotter, 1971; Baker, 1973a; Padgett and Ehrlich, 1976). The Manning equation is expressed by the relationship:

$$V = 1.49R^{2/3}S^{1/2}/n \tag{16}$$

where V is velocity in ft s^{-1}, S is slope, R is hydraulic radius, and n is Manning's roughness coefficient. For streams with large width to depth ratios, such as braided streams, the hydraulic radius is approximately equal to stream depth (Baker, 1973b; Bradley and Mears, 1980). Manning's equation can then be expressed as:

$$D = \left(Vn/1.49S^{1/2}\right)^{3/2} \tag{17}$$

Bray (1979) presented several methods for evaluating Manning's n. Two of these methods, by Cowens and by Strickler, were used to determine a range of n values for this study. Cowens' method (in Bray, 1979) is expressed by the equation:

$$n = \left(n_0 + n_2\right)M_c \tag{18}$$

where n is Manning's roughness coefficient, n_0 is roughness related to median bed material size and is taken as 0.036 for particles between 64 and 128 mm, M_c is a meandering factor equal to 1.15 since the Ballena river sinuosity was greater than 1.2, and n_2 is an additional roughness component related to variations in channel cross-section and is taken as 0.005 for this study. Based on these values, an estimation of n from the Cowens method is 0.047. The Strickler method (in Bray, 1979) for determining Manning's n is based directly on particle size and is given by the equations:

$$n = 0.041D_{50}{}^{1/6} \tag{19}$$

and:

$$n = 0.038D_{90}{}^{1/6} \tag{20}$$

where D_{50} and D_{90} are characteristic bed material sizes given in m. From the cumulative curves D_{50} is taken as 0.06 m which produces an n value of 0.026, and D_{90} is 0.25 m yielding an n value of 0.030. Based on these two methods, the extreme n values of 0.026 and 0.047 along with the previously determined values of V and S are used to evaluate Manning's equation.

Lacey (in Bray, 1979) presented an empirical equation that is similar to the Manning equation; however, roughness is implicit and does not have to be evaluated independently. The Lacey equation is:

$$V = 10.8d^{2/3}S^{1/3} \tag{21}$$

where V is mean velocity in m s^{-1}, d is mean water depth in ft, and S is average slope of the water surface. Solving eq. 21 for d gives:

$$d = \left(V/10.8S^{1/3}\right)^{3/2} \tag{22}$$

Another equation used to estimate paleodepth is the DuBoys equation for boundary shear (Bradley et al., 1972; Baker, 1973; Koster, 1978; Bradley and Mears, 1980). The equation is:

$$T_c = \gamma RS \tag{23}$$

where T_c is critical shear stress in lb ft^{-2}, γ is specific weight of transporting fluid, S is slope and R is hydraulic radius in ft. As before, flow depth is substituted for R and the equation is rearranged in terms of D to give:

$$D = T_c/\gamma S \tag{24}$$

where D is average flow depth in ft. γ is taken as 1000 kg m^{-3} for the density of clear water and as 1070 kg m^{-3} for water with 10% suspended sediment by weight (Bradley and Mears, 1980).

A similar method was developed by Baker (1974) based on the shear stress relations of Shields. The equation, solved for depth (d) in m, is:

$$d = 0.06D(\gamma_p - \gamma_f)/Sr_f \tag{25}$$

where D is competent particle size in m, γ_p and γ_f are specific weights of particle and fluid, respectively, and S is slope. For rhyolites, γ_p is taken as 2520 kg m^{-3} (Telford et al., 1978) and γ_f is the same as in eq. 24. Values of S range from 0.0025 to 0.0083 as determined previously. For this method, dimensionless shear stress (Baker, 1974) is assumed to stay constant at a value of 0.06 for particles larger than 0.7 cm.

Koster (1978) published an equation for estimation of paleodepth based on the DuBoys shear stress equation, particle and fluid densities, and particle size. The equation is designed for paleodepth estimations in relatively shallow, gravelly, braided rivers. Koster's equation is:

$$D = 0.15d^{0.67}/S \tag{26}$$

where D is average depth in cm, d is competent particle size in cm (52), and S is slope.

Most engineering equations involving depth usually relate either velocity or discharge as a function of depth. However, a method designed for gravel streams by Parker (1979) relates depth to slope by the equation:

$$H^* = 0.0553S^{-1.013} \tag{27}$$

where H^* is depth defined as D/d_{50} where D is the average depth, d_{50} is mean particle size, and S is slope. Any consistent set of units may be used in the equation. Solving eq. 27 for D gives:

$$D = 0.0553S^{-1.013}d_{50} \tag{28}$$

Equations 24 and 28 are closely related because they have been derived in part from shear stress relations.

Cross-stratified sandstone lenses are uncommon in the Ballena Gravels. Due to the extremely coarse sediment population, bars are composed mainly of gravels that rest upon gravels. However, at several localities sandstone lenses are found up to 2 m thick. They were most likely deposited during waning flood stages when water depths were low; thus, they may not be indicative of bankfull conditions.

Bankfull discharge is defined by workers in different ways. It is usually equated with a flood that has a recurrence interval of 2.33 yrs. However, Kellerhals' (1967) techniques give estimates of discharges with a return period of 5 to 20 yrs.

Results of depth calculations. Depth values obtained from the seven techniques are tabulated in Table III and presented as average depths for bankfull conditions. The four values of shear stress used in the DuBoys equation are shown in Table III and average values for the four computations are given.

The DuBoys equation produces the largest (55 m) as well as the broadest range of depth values. The lowest paleodepth values are from the Manning (0.6 m) and Parker (0.45 m) techniques. Low values from the Manning equation are produced when velocities and roughness are low (2.5 m s^{-1}, 0.026) and when the channel slope is high (0.0083).

The highly variable values generated from the DuBoys equation are in part related to scatter in the initial shear stress data plots. Velocity values determined from the same shear stress values have much less variation. Bradley and Mears (1980) suggested that this is expected since depth and tractive force vary proportionately but velocity changes more slowly than tractive force. Since slope values are already relatively low, the extreme values of depth determined from the DuBoys equation must be directly related to the shear stress variable.

Inspection of Table III shows that six of the seven techniques overlap within the narrow range from 2.5–3 m. Significant overlap in depth values also occurs in the

TABLE III

Depth results

	T_c (kg dm^{-2})	Bankfull depth (m)
Manning equation		0.6 to 7.2
Lacey equation		1.1 to 4.5
DuBoys equation		
Baker (1974)	8.2 to 185	1.1 to 88
Shields (1936)	42.1	5.6 to 19.9
Fahnestock (1963)	12.4 to 41.2	1.6 to 19.5
Church (1972)	12.8 to 193	1.7 to 91.9
Mean	19.0 to 116	2.5 to 54.8
Bradley and Mears (1980)		1.6 to 5.8
Koster (1978)		2.6 to 8.5
Parker (1979)		0.4 to 1.4
Sand lenses		0.6 to 3.0

range of 2.5–4.5 m where five of the seven techniques are in agreement. From the multiple techniques approach it is concluded that average water depths during bankfull conditions for the ancient Ballena River were probably in the range of 2.5–4.5 m.

Discharge

Several different methods have been used to estimate discharge in ancient streams. These methods are based on empirical data and are related to variables such as width, depth, gradient, sediment characteristics, bed roughness, meander wavelength, river length, area of drainage basin, and velocity. The simplest equation for discharge is:

$$Q = VA \tag{29}$$

where Q is discharge, V is mean velocity, and A is cross-sectional area. This equation is commonly used to estimate discharge in both modern and ancient rivers where values of V and A can be determined. However, since the channel width necessary for determining A in this study is difficult to determine, other means of estimating discharge are used.

Although most studies have been concerned with sand-dominated fluvial systems, several engineering studies have been conducted on gravel-dominated rivers to aid in the design of artificial channel systems (Kellerhals, 1967; Parker, 1979; Chang, 1980). In general, the equations relate discharge to channel morphology. As many of the data are based on studies of natural fluvial channels, they were used also for paleohydrologic determinations in this study.

Criteria used in the engineering studies are that: (1) the channel is straight and symmetrical, (2) most grain sizes are mobile at bankfull discharge, (3) the channel bed is composed of gravel, (4) the channel perimeter is composed of either loose or immobile gravel depending on the study, and (5) the streams are in regime or quasi-equilibrium. For paleohydrologic studies, only a rough estimate of flow conditions is obtained and these conditions can be assumed to have been met.

Six techniques were used to estimate paleodischarge for the Ballena river. Three of the techniques are engineering equations relating discharge to slope, two other engineering equations relate discharge to depth, and one technique relates stream length to drainage area and discharge.

Discharge as a function of gradient. Parker (1979), Kellerhals (1967), and Chang (1980) have developed equations relating discharge to gradient based on studies of gravel-bearing systems. The equations used were developed semi-independently of one another since there was some overlap in the acquired data.

Parker (1979) published the equation:

$$S = 0.223\tilde{Q}^{-0.410} \tag{30}$$

where S is slope and $\tilde{Q}$ is dimensionless discharge expressed by:

$$\tilde{Q} = Q/(RgD_{50})^{0.5}D_{50}{}^{2} \tag{31}$$

where D_{50} is median grain diameter, R is given as $(\rho_s/\rho_w - 1)$, ρ_s is density of sediment taken as 2.52 and ρ_w is water density, g is gravitational acceleration, and Q is bankfull discharge. As before, any consistent set of units can be used to evaluate Parker's equation. When values of g, R, D_{50} are substituted into eq. 31, it can be solved in terms of discharge (Q) to give:

$$Q = 0.0031S^{-2.44} \tag{32}$$

where Q is bankfull discharge in $ft^3\ s^{-1}$.

Kellerhals (1967) analyzed data from rivers, canals, and flumes that were transporting coarse sediment. He used three independent relations—width vs. discharge, a bedload equation, and a tractive force equation to solve for the regime equation:

$$S = 0.120Q_d^{-0.4}k_s^{0.92} \tag{33}$$

where S is slope, Q_d is dominant or channel-forming discharge in $ft^3\ s^{-1}$, and k_s is bed roughness approximately equal to D_{10} in ft. Substituting for k_s and solving in terms of discharge, eq. 33 becomes:

$$Q_d = 0.0032S^{-2.5} \tag{34}$$

Chang (1980) developed a similar equation for gravel-bearing systems based upon a resistance equation, a bedload equation, and the condition of minimum stream power. Chang showed that discharge is proportional to the sediment moving through the channel and the channel gradient. Chang's equation for active gravel streams is:

$$S = 0.001065d_{50}{}^{1.15}Q^{-0.42} \tag{35}$$

where S is channel slope, Q is bankfull discharge in $ft^3\ s^{-1}$, and d_{50} is median grain size in mm. Equation 35 solved for Q becomes:

$$Q = 0.0062S^{-2.38} \tag{36}$$

Discharge as a function of depth. Discharge is also commonly expressed as a function of channel depth and bed-material size. Regime formulas, similar to the slope–discharge equations, have been developed for coarse-grained river systems. They are based on empirical data obtained from canals, laboratory flumes, and natural rivers.

Kellerhals (1967) determined the following regime equation for mean flow depth as a function of discharge and bed roughness:

$$D = 0.166Q^{0.4}k_s^{-0.12} \tag{37}$$

where D is mean flow depth in ft, Q is bankfull discharge in $ft^3\ s^{-1}$, and k_s is grain roughness equivalent to mean grain size in ft. If k_s is set equal to 0.820 (25 cm) and the equation solved for Q, then eq. 37 becomes:

$$Q = 83.9D^{2.5} \tag{38}$$

Parker (1979) arrived at a similar equation:

$$H_c^* = 0.253\tilde{Q}^{0.415} \tag{39}$$

where H_c^* is the dimensionless depth of the central region of the channel given by H_c/d_{50}, where d_{50} is median particle size, H_c is depth in the central region, and $\tilde{Q}$ is dimensionless discharge as in eq. 31. Solving for Q, eq. 39 becomes:

$$Q = 165.9H_c^{2.41} \tag{40}$$

where H_c is in ft and Q is in ft^3 s^{-1}.

Discharge based on stream length. Another independent means of estimating paleodischarge is based on the relations between stream length and drainage basin area and between drainage basin area and discharge.

Several workers have shown that the area of a drainage basin is related to the discharge of the main channel from that drainage basin (Leopold et al., 1964; Hack, 1957). Leopold et al. (1964) showed that bankfull discharge and drainage basin area are related by the equation:

$$Q \propto A_d^{\,n} \tag{41}$$

where Q is bankfull discharge in ft^3 s^{-1}, A_d is drainage basin area in miles^2, and n is a positive exponent. They showed that the exponent (n) is controlled by climate and varies from 0.65 to 1, and that 0.75 is average. In humid regions, the exponent for discharge is closer to 1, whereas arid regions would have n-values closer to 0.65 (Hack, 1957; Leopold et al., 1964; Eicher, 1969). During the late Eocene, the San Diego area is considered to have had a semi-arid climate (Peterson and Abbott, 1979). This suggests that the exponent for eq. 41 was probably low, around 0.65. However, n-values of both 0.65 and 1 are used herein to estimate Q.

To establish paleodischarge by this method, an estimate of the drainage basin area for the Ballena river is necessary. Hack (1957) measured 90 drainage basins in Virginia and Maryland and Langbein et al. (1949) measured about 400 basins in the northeastern United States. These studies showed that length of stream channels and area of drainage basins were strongly related by the equation:

$$L = 1.4A_d^{\,0.6} \tag{42}$$

where L is channel length in miles, and A_d is drainage basin area in miles^2. Rearranging eq. 42 in terms of drainage area (A_d) gives:

$$A_d = 0.57L^{1.67} \tag{43}$$

Earlier it was shown that the probable length of the Ballena channel was between 250 and 400 km. If these values of channel length are substituted into eq. 43, it suggests that the drainage area for the Ballena Gravels was between 6500 and 14,000 km^2.

Leopold et al. (1964) compiled additional information for various drainage basins of the world. Their data include information from many large rivers, whereas Hack's data are generally restricted to streams less than 80 km in length. When two parallel lines are drawn incorporating the data of Leopold et al. (1964), a range of values for drainage area based on channel length is produced. Values of drainage area from this method are somewhat greater than those values produced by Hack's method and range from 5200 to 26,000 km^2.

A drainage basin area estimate also can be made based on the Eocene paleogeography. Eocene fluvial gravels that are coeval with the Ballena Gravels occur in the Valle de Las Palmas area of northern Baja California (Fig. 1). These gravels which lie approximately 90 km to the south of the Ballena channel also trend in a similar direction and may have originated from a related source terrane (Minch, 1972). No other Eocene conglomerates of fluvial or any other origin are present between the two systems.

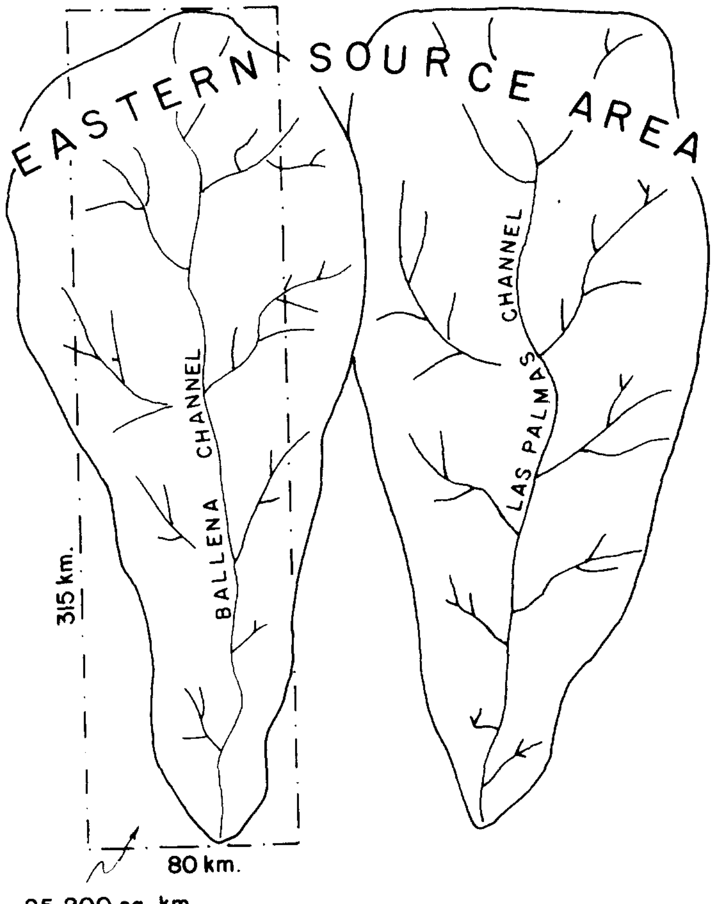

Fig. 8. Drainage area estimates based on paleogeography for the Ballena river and the Las Palmas river, the next adjacent Eocene drainage to the south.

Using the Valle de Las Palmas channel as a guide and assuming that the drainage area between the two systems was equally divided, an estimate of the drainage area for the Ballena channel can be made. Figure 8 shows that based on a channel length of 315 km and an average drainage basin width of 80 km, the drainage area would have been 25,200 km^2.

Finally, substituting the values of drainage basin area into eq. 41 produces a range of discharge values that are tabulated in the following section.

Results of discharge calculations. Results of the six different techniques used to estimate paleodischarge are tabulated in Table IV. The discharge values are given as a range of values since the various techniques are themselves based upon ranges of gradient, depth, and stream length. The discharge values represent bankfull conditions.

The greatest values of discharge were produced from the two techniques based on channel depth with Parker's technique yielding the greatest value. Both the slope and stream length–drainage area techniques produce lower, less variable discharge results. The extreme values of discharge are produced when water depths are great (4.5 m) and gradients are low (0.0025). Discharge values determined from the lower end of the depth range (2.5 m) are much closer to values generated from the gradient and length–area techniques.

The large variability found within the discharge results may be related to several factors. First, the engineering equations used in the study are highly sensitive. Small

TABLE IV

Results of discharge calculations (in $cm^3\ s^{-1}$)

Discharge based on:					
Gradient		*0.0025*		*0.0083*	
Parker (1979)		200		11	
Kellerhals (1967)		290		15	
Chang (1980)		274		16	
Depth	*4.5 m*		*2.5 m*		*1.5 m*
Parker (1979)	3210		735		225
Kellerhals (1967)	2070		430		135
Stream length – drainage area					
	n = 1, maximum area			*n = 0.65, minimum area*	
Hack (1957)	155			to 6	
Leopold et al. (1964)	283			to 4	
Paleogeographic reconstruction		*n = 1*	*n = 0.65*		
		276	to 43		

changes in either depth or slope produce relatively large changes in discharge values (e.g., decreasing the lower value of slope, 0.0025, by half raises discharge values from approximately 283 $m^3\ s^{-1}$ up into the 1420 $m^3\ s^{-1}$ range).

The extremely large values of discharge are directly related to the estimated depth value of 4.5 m, which in part was calculated to transport the coarsest fraction of the sediment load and would have existed only when the Ballena river was at a high flood stage. It should be noted that discharge estimates produced from the six techniques only represent bankfull conditions; large floods of rare occurrence are not taken into account. Water depths during bankfull conditions were probably less than those estimated from the competent velocity–slope calculations. This also suggests that bankfull discharges were less than those determined by the large depth values used in the computations. If water depths of 1.5 m are used in the equations for bankfull depths, values drop to 135–225 $m^3\ s^{-1}$ (Table IV). These values are in closer agreement with those generated from the other techniques.

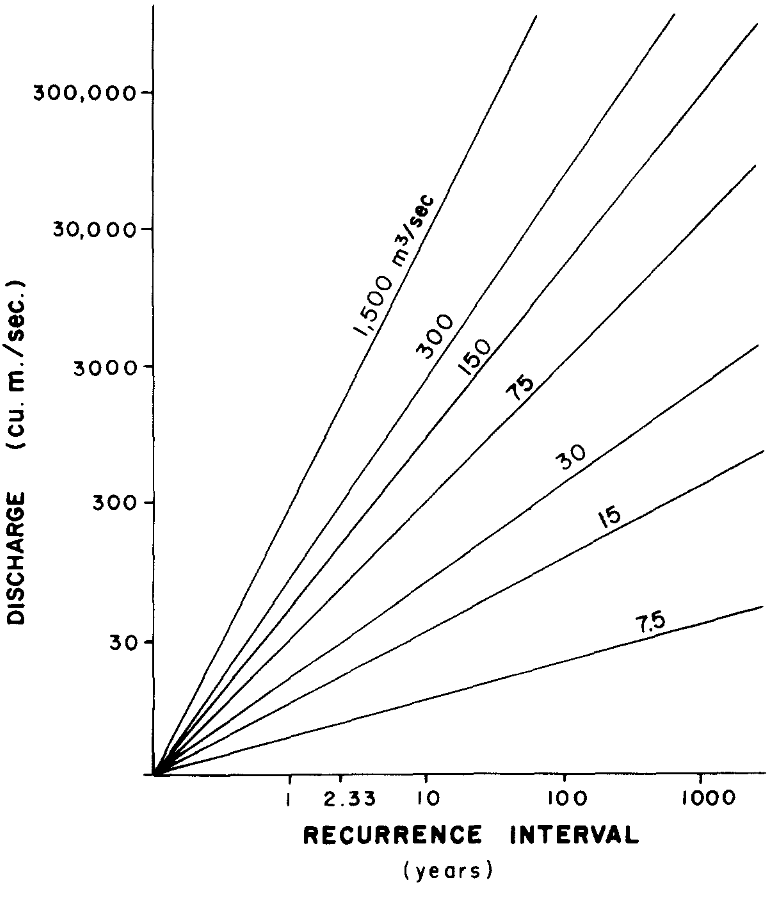

Fig. 9. Flood magnitude and recurrence based on different values of mean annual flood.

If it is assumed that the estimated discharge values are approximately equal to the 2.33 yr flood, then the magnitude of major floods on the Ballena river can be estimated. It has been shown that the frequency and magnitude of floods commonly have a logarithmic nature (Leopold et al., 1964; Morisawa, 1968). Figure 9 gives possible discharge values for floods of varying frequency based on a logarithmic frequency–magnitude curve and differing values of flood recurrence intervals. Figure 9 shows that if the 2.33 yr flood is approximately 275 $m^3 s^{-1}$, discharges on the order of 2750 $m^3 s^{-1}$ would be expected for the 10 yr flood and discharges up to 27,500 $m^3 s^{-1}$ might occur once in 100 yrs.

Width

One of the strongest empirical relations in fluvial geomorphology is that between channel width (B) and bankfull discharge (Q). In this section, channel widths are estimated using this relationship and the previously determined values of discharge.

Equations developed by Kellerhals (1967), Parker (1979), and Chang (1980) relating channel width to discharge are used to estimate ancient channel widths of the Ballena river. The engineering equations are based on data acquired from studies of flumes, canals, and natural gravel bed streams.

Kellerhals (1967) determined an entirely empirical relation for stream width given by:

$$B = 1.8Q^{0.5} \tag{44}$$

where B is water surface width of the channel in ft, and Q is dominant (bankfull) discharge in $ft^3 s^{-1}$.

Parker (1979) derived a similar empirical equation:

$$B^* = 4.4\tilde{Q}^{0.5} \tag{45}$$

where $\tilde{Q}$ is dimensionless water discharge as defined previously for eq. 36, and B^* is the dimensionless channel width given by B/D_{50}, where B is channel width and D_{50} is median grain size. Since Parker's equation involves dimensionless parameters, they can be used with any consistent system of units. Equation 45 is simplified to give:

$$B = 2.5Q^{0.5} \tag{46}$$

where B is channel width in ft and Q is bankfull discharge in $ft^3 s^{-1}$. Parker (1979) stated that the data plot coherently with little scatter but also added that a good "rule" is to determine empirically the relationship, in the form of eq. 45, for each watershed. However, Parker also noted the similarities of eq. 45 with relationships developed by other workers such as Leopold and Maddock (1953) and suggested that some universality may exist within the relation.

Chang (1980) determined the following regime equation for channel width:

$$B = \left\{1.905 + 0.249\left[\ln\left(0.001065{d_{50}}^{1.15}/SQ^{0.42}\right)\right]^2\right\}Q^{0.47} \tag{47}$$

where B is channel width in ft, d_{50} is median grain size in mm, S is slope, and Q is bankfull discharge in ft^3 s^{-1}. Chang's equation is not entirely empirical as were the preceding two techniques, but is derived from several established stream relationships. Chang (1980) noted that the analytical channel width determined from the equation is proportional to $Q^{0.47}$ and that except for steep slopes, where the width increases rapidly with gradient, the width is only a function of discharge.

Results of channel width calculations. Calculations of channel width from the three techniques are tabulated in Table V. The results are all in general agreement. The closeness of the results may be related to a strong correlation between the discharge–width variables as indicated by the similarities in the major exponents of the equations. Also, a degree of overlap occurs in the data collected by the three workers.

Many paleohydrologic studies had channel width measured in the field directly from the rocks. However, for the Ballena Gravels, the lack of outcrop continuity and the poor bedding characteristics of the rocks make this simplest of techniques difficult to use, if not impossible.

Assuming the calculated discharge values are approximately correct, the ancient Ballena river probably had 1 to 2.33 yr bankfull channel widths ranging between 23 and 76 m.

Rainfall

Schumm (1968) showed that mean annual precipitation can be estimated by expressing rainfall as a function of mean annual runoff and temperature (Fig. 10). He recognized that the data determined by Langbein et al. (1949) were applicable to many different climatic conditions. In order to determine ancient precipitation values for the Ballena river based on Fig. 10, estimations of runoff and temperature are necessary.

Runoff can be estimated if it is assumed that water flowing off the land will eventually find its way through the drainage net into the main channel. Based on this assumption, runoff can be set approximately equal to discharge, and mean

TABLE V

Results of width calculations

Techniques		Discharge	
		71 m^3 s^{-1}	283 m^3 s^{-1}
Kellerhals (1967)		27 m wide	55 m wide
Parker (1979)		38 m	76 m
Chang (1980)	gradient =0.0025	23 m	44 m
	0.0083	24 m	52 m

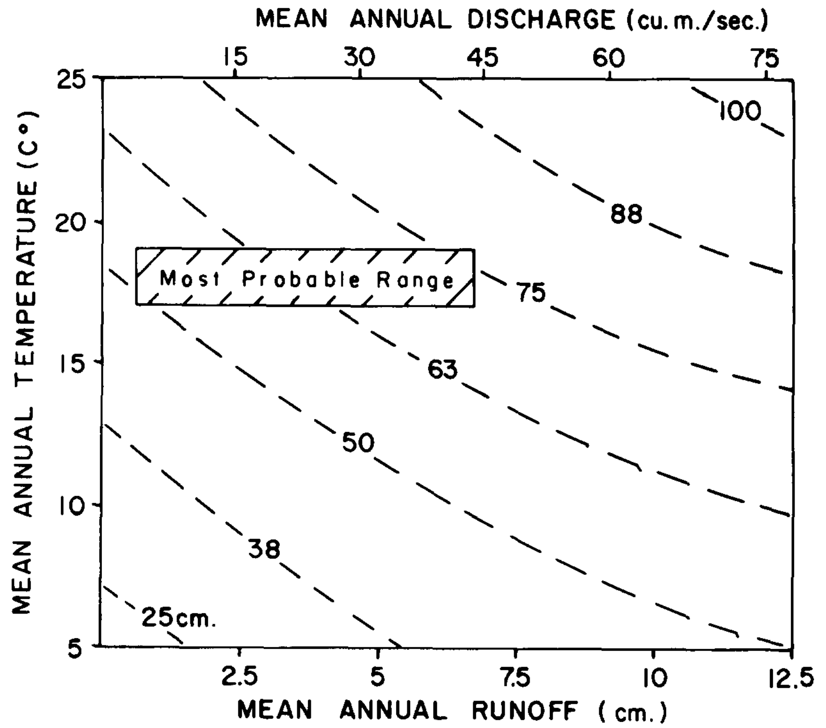

Fig. 10. Rainfall curves for the Ballena river drainage area based on mean annual temperature, mean annual discharge, and mean annual runoff.

annual runoff will be approximately equal to mean annual discharge. The bankfull discharge values calculated earlier for the Ballena river need to be converted to mean annual discharge.

Several studies have related both mean annual flood, which is approximately equal to bankfull discharge, and mean annual discharge to flow parameters that are equivalent. If the two relationships are equated by a like variable, then mean annual discharge can be expressed as a function of bankfull discharge. Three different pairs of equations used to determine the conversion are presented below.

In the first method, Dury (1965) and Carlston (1965) arrived at relationships expressing the two discharges as a function of meander wavelength. Dury's (1965) equation relating meander wavelength to mean annual flood is given by:

$$\lambda = 30Q_{\mathrm{ma}}^{0.5} \tag{48}$$

where λ is a meander wavelength and Q_{ma} is mean annual flood. Carlston (1965) arrived at a similar relationship; however, meander wavelength was expressed as a function of mean annual discharge and is expressed as:

$$\lambda = 106Q_{\mathrm{m}}^{0.46} \tag{49}$$

where Q_m is mean annual discharge. Equations 48 and 49 are set equal as follows:

$$30Q_{ma}^{0.5} = 106Q_m^{0.46} \tag{50}$$

By simplifying eq. 50, the following expression giving mean annual discharge as a function of mean annual flood was obtained:

$$Q_m = 0.064Q_{ma}^{1.087} \tag{51}$$

Schumm (1967) presented two equations, both related to meander wavelength, that can be equated as above:

$$L = 1890Q_m^{0.34}/M^{0.74} \tag{52}$$

and:

$$L = 234Q_{ma}^{0.48}/M^{0.74} \tag{53}$$

where L is meander wavelength and M is Schumm's sediment load parameter (Schumm, 1963). Equating and solving the two equations for Q_m gives:

$$Q_m = 0.0021Q_{ma}^{1.41} \tag{54}$$

A final method, also using equations supplied by Schumm (1968), expresses channel depth as a function of both mean and bankfull discharge. The equations are given by the following:

$$d_b = 0.6M^{0.34}Q_m^{0.29} \tag{55}$$

and:

$$d_b = 0.09M^{0.35}Q_{ma}^{0.42} \tag{56}$$

and solving for Q_m gives:

$$Q_m = 0.00144Q_{ma}^{1.45} \tag{57}$$

When previously determined values of bankfull discharge, which are taken as 71–283 m^3 s^{-1}, are substituted into eqs. 51, 54, and 57, values for mean annual discharge ranging from 3 to 40 m^3 s^{-1} are produced.

The data of Schumm, Bray, and Carlston were collected on sand-dominated, perennial streams that probably have a somewhat higher mean annual discharge than intermittent, braided streams. This suggests that the values of mean annual discharge determined from the equations represent maximum values and that the true mean annual discharge might be toward the lower end of the determined range.

Runoff is determined by converting the values of mean annual discharge determined above to annual discharge and dividing by the drainage area. Annual discharge is the amount of water flowing through the channel for a year and the calculated values range from 1.1×10^8 to 1.3×10^9 m^3. These values are assumed to be approximately equal to annual runoff in m^3. A value for drainage area, which was previously determined, is 25,000 km^2 or approximately 2.5×10^{10} m^2. Dividing the annual runoff values by watershed area produces mean annual runoff values from 0.44 to 5.2 cm.

An estimate of paleotemperature was made by Peterson and Abbott (1979) from paleoclimatic indicators within the Poway Group sedimentary rocks and coeval Ballena Gravels. They interpreted caliche horizons, which are up to 1 m thick, to represent soil horizons that formed where mean annual temperature was great enough to produce strongly evaporative conditions. They interpreted in situ, fractured clasts as having been split by salt weathering, and the immature clay-mineral suite dominated by smectite and vermiculite to represent weathering in a semi-arid climate. They concluded that the average annual temperature for the Late Eocene in the San Diego area was probably from 18 to 20°C.

Based on these estimates of runoff and temperature, an estimate of precipitation can be read from Fig. 10. Values corresponding to the "most probable range" are enclosed by the calculated runoff values and the narrow range of paleotemperatures determined by Peterson and Abbott (1979). The values range from 50 to 75 cm of precipitation annually.

These findings, which are based almost entirely on quantitative methods, are in close agreement with the 51 to 64 cm of annual rainfall estimated by Peterson and Abbott (1979) for the Late Eocene in the San Diego area. It should be noted that these precipitation estimates are based on calculations and assumptions that are subject to considerable error. Therefore, a large degree of confidence should not be placed on the results.

MODERN ANALOGUES TO THE BALLENA RIVER

Nueces River

The Nueces River of Texas has many characteristics that are similar to the Ballena river (Table VI). The Nueces originates in the Edwards Plateau, is approximately the same length, and has similarities in sediment size, drainage area, and climate (Gustavson, 1978). The Nueces River has a highly varied discharge which also is interpreted to be characteristic of the Ballena river. Maximum flood discharge on the Nueces was 17,460 m^3 s^{-1} on June 11, 1935 after a storm dumped 15 cm of rain on most of the drainage basin. Gustavson (1978) suggested that floods of this magnitude on the Nueces have a return period on the order of 50 yrs. Flood magnitude-frequency estimates suggest that values of this magnitude are reasonable for the Ballena river (Fig. 9).

Velocities measured on the Nueces River range from 2.7 to 4.4 m^3 s^{-1} for floods when most or all of the sediment load was active. These values coincide closely with estimates for the ancient Ballena river. Flood velocities up to 6 m s^{-1} have been recorded on the Nueces, whereas three of the techniques used to estimate velocity on the Ballena river produced values in excess of 6 m s^{-1}. The Ballena river apparently was affected by heavy rainfalls during short spans of time as characterize the Nueces

TABLE VI

Comparison of ancient Ballena river to the Nueces River of Texas

Characteristics	Ballena river	Nueces River (Texas) (from Gustavson, 1978)
Sediment size	Coarse gravel	Coarse gravel
Mean grain size	3.7 to 5.6 cm	1.2 to 3.4 cm
Large clast size	> 25 cm	> 11 cm
Lithology	Meta-volcanic	Limestone
Stream length	About 315 km	484 km
Drainage area	About 25,000 km^2	About 31,000 km^2
Sinuosity	1.1 to 1.3	1.3
Slope	7 to 21 m km^{-1}	About 2 m km^{-1}
Elevation of source area	2150 to 4000 m	600 to 900 m
Discharge:		
Maximum	> 14,000 m^3 s^{-1}?	17,460 m^3 s^{-1}
2-yr flood	6 to 275 m^3 s^{-1}	340 m^3 s^{-1}
Mean	3 to 40 m^3 s^{-1}	3 m^3 s^{-1}
Minimum	0?	0
Velocities	2.5 to 4 m s^{-1}	2.7 to 4.4 m s^{-1}
Maximum	7.5 m s^{-1}?	6 m s^{-1}
Flow conditions	Highly variable	Highly variable
Flow regimes	Upper and lower	Upper and lower
Climate	Semi-arid	Semi-arid
Precipitation	49 to 79 cm	51 to 56 cm
Temperature	18 to 20°C	21°C

and other rivers that cross the Balcones escarpment of the Edwards Plateau in south-central Texas today.

The major discrepancy found between the two rivers is the elevation of the source area. The Nueces River source is much lower than the elevations estimated for the Ballena river source. This elevation factor influences channel slope, which is lower for the Nueces. Nonetheless, it appears that the Eocene Ballena river had some similarities with the modern Nueces River of Texas.

Southern Argentina rivers

Rivers in southern Argentina have characteristics similar to those estimated for the Ballena river (Table VII). Four rivers, the Rio Colorado, Rio Negro, Rio Chubut, and Rio Santa Cruz flow down the eastern flank of the Andes Mountains and eastward into the Atlantic Ocean. The rivers originate in snow-covered mountains at elevations from 2150 to 3350 m and stream lengths range from 400 to 855 km. Since the Argentine rivers are supplied in part by Andean meltwaters, their discharges are not as variable as those estimated for the Ballena river. The Argentine rivers have higher mean and minimum discharges and lower maximum discharges. By contrast,

TABLE VII

Comparison of ancient Ballena river to rivers of southern Argentina

Characteristics	Ballena river	Argentine rivers (data from Van der Leeden, 1975)			
		Rio Colorado	Rio Negro	Rio Chubut	Rio Santa Cruz
Length	250 to 400 km	855 km	645 km	484 km	400 km
Elevation of source area	2150 to 4000 m	3050 m	2500 m	2150 m	3350 m
Drainage area	25,000 km^2	22,360 km^2	93,600 km^2	15,600 km^2	15,600 km^2
Slope in lower reaches	7 to 21 m km^{-1}	1.5 m km^{-1}	1.5 m km^{-1}	2 m km^{-1}	5 m km^{-1}
Sediment size	Gravel	Gravel	Gravel?	Gravel?	Gravel
Discharge					
Maximum	27,500 $m^3 s^{-1}$?	822 $m^3 s^{-1}$	3400 $m^3 s^{-1}$	540 $m^3 s^{-1}$	2000 $m^3 s^{-1}$
Minimum	0 $m^3 s^{-1}$?	31 $m^3 s^{-1}$	74 $m^3 s^{-1}$	4 $m^3 s^{-1}$	195 $m^3 s^{-1}$
Mean	3 to 40 $m^3 s^{-1}$	130 $m^3 s^{-1}$	935 $m^3 s^{-1}$	48 $m^3 s^{-1}$	680 $m^3 s^{-1}$
Climate					
Lower reaches	Semi-arid	Semi-arid	Semi-arid	Semi-arid	Semi-arid
Headwaters	Mountains	Snow-covered mountains	Snow-covered mountains	Snow-covered mountains	Snow-covered mountains

the Ballena river evidently was dry much of the year but flooded seasonally following major storms.

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