

Shoreline Processes and the Age and Elevation of the Lake Lahontan Highstand in the Jessup Embayment, NV.

Kenneth D. Adams
Steven G. Wesnousky
Center for Neotectonic Studies and
Department of Geological Sciences
University of Nevada Reno

Abstract

Shoreline deposits and landforms in the Jessup Embayment are used to refine the timing and magnitude of the last highstand of pluvial Lake Lahontan. Modern analogs of coarse clastic barrier systems are used to interpret both form and process of coarse clastic paleo-beach barrier features. Shore features are broadly classified into erosional and constructional forms and related to 1) preexisting morphology of the shoreline and lake bed 2) sources, physical properties and rates of sediment supply 3) climatic and wave energy environment and 4) the rate and direction of lake level change. Gastropod shells collected from an exposure along a progradational barrier complex near the head of the Embayment yielded AMS ^{14}C ages of $13,280 \pm 110$ and $13,110 \pm 110$ yrs BP and define the lake level during transgression when the surface of the Lake was at about 1326 and 1330 m, respectively. An AMS ^{14}C age of $12,690 \pm 60$ yrs BP for a camel bone found behind a highstand barrier in a paleo-lagoon places a minimum limiting age on the highstand of Lake Lahontan which reached an elevation between 1338 and 1339 m in the Jessup Embayment. The maximum age of the highstand is constrained by the youngest age of the transgressive deposits (~ 13.1 ka) which predate the highstand. The elevation (~ 1338.5 m) and timing ($\sim 12.7 - 13.1$ ka) of the highstand imply that its magnitude was larger and its age younger than previous estimates. During the regression from the highstand, the Lake formed at least 28 barriers which are used to delineate the relative rate and character of the regression.

Introduction

The shoreline deposits and landforms of pluvial Lake Lahontan display both a spectacular proxy record of climate change and an excellent sedimentary record of shoreline processes operating on coarse clastic systems. The highest shoreline of Lake Lahontan prescribes an extremely complex path of some 3000 km, dictated by Basin and Range topography, and has many embayments, headlands and islands (Figure 1). It is the preservation of the Lahontan shore record which supplies evidence for the number and timing of lake cycles, the levels attained by these various lake cycles through time, and the coastal processes and paleoenvironmental interpretations made from these deposits and landforms.

The Jessup Embayment is a small bay located in the northwestern part of the Carson Sink (Figure 1) that was inundated during the last major highstand of Lake Lahontan. The Embayment opens to the southeast and is fronted by an arm of the Carson Sink playa which lies at an elevation of about 1185 m. The highest constructional shorelines in the Jessup Embayment are found at about 1340 m; therefore the shorelines record vertical water fluctuations of about 155 m. Shoreline features in the Embayment are particularly well-developed because of the large fetch (~ 60 km) to the southeast (Figure 1) and provide a detailed record of fluctuations and shore processes operating during both transgressive and regressive stages of the last lake cycle.

The purpose of this paper is to elucidate the lacustrine history of the Jessup Embayment utilizing geomorphic, sedimentologic, stratigraphic and age dating techniques. We present a new AMS radiocarbon date on a camel bone found in a lagoon behind a high barrier that provides a closely limiting age on the last highstand of Lake Lahontan. The observations and results of modern process studies of coarse clastic barrier systems are used as modern analogs to better interpret the landforms and deposits created by pluvial Lake Lahontan. Particular attention is paid to features and deposits of the last major lake cycle because they constitute the vast majority of surfaces and exposures.

Previous Work

Israel Russell (1885) performed the seminal work on Lake Lahontan and to this day his monograph remains the only comprehensive study which covers the entire Basin. From exposures within the major river canyons, Russell (1885) recognized the deposits of two major lake cycles separated by a package of subaerial sediments which he named the medial gravels. He termed the lacustrine deposits the upper and lower lacustral clays, which represent the last and penultimate lake cycles, respectively. Morrison (1964, 1991) redefined Russell's stratigraphy and named the deposits of the last major lake cycle the Seho AF and deposits of the penultimate lake cycle the Eetza AF. Subaerial and shallow lake sediments deposited between the Eetza and Seho lake cycles are now known as the Wyemaha AF. Morrison (1991) claims that the Eetza AF may represent as many as eight lake cycles which are thought to be broadly correlative with marine oxygen isotope stages 6, 8 and 10 (130-360 ka). However, age control on the Eetza AF is limited to two tephra exposures (Wadsworth tephra) along the Truckee River near Wadsworth, NV (Davis, 1978; Morrison, 1991) and several uranium-thorium dates ranging from 128 to > 300 ka on gastropod shells (Kaufman and Broecker, 1965), all of which are located within lacustrine deposits below an elevation of 1265 m. The age of the Wadsworth tephra is not agreed upon but probably lies between about 150 to 200 ka (Berger, 1991; Morrison, 1991). Shoreline deposits from the Eetza AF above 1265 m have never been directly dated and age assignments appear to be based on correlating coarse clastic shore gravels with fine grained inner basin deposits.

The age of the Seho AF is much better constrained than earlier lake cycles with literally hundreds of radiocarbon dates on tufa, gastropods, wood, bones and other carbon-bearing material (Broecker and Kaufman, 1958, 1965; Kaufman and Broecker, 1965; Born, 1972; Davis, 1978, 1982, 1983; Benson, 1978, 1981, 1991, 1993; Benson and Thompson, 1987a, 1987b; Benson et al, 1990, 1992, 1995; Benson and Peterman, 1995; Thompson et al, 1986, 1990; Lao and Benson, 1988; Dansie et al, 1988; Dorn et al, 1990). In addition, Davis (1978) has identified no less than 11 individual tephra horizons within the Seho AF which have been instrumental in determining its age as well as correlating isolated stratigraphic sections. The beginning of the Seho period is debatable, but is commonly thought of as being correlative with oxygen isotope stage 4 (~70 ka) (Lao and Benson, 1988; Dansie et al, 1988). However, Morrison (1991) has proposed that the base of the Seho be placed at the time of deposition of the Marble Bluff (Saint Helens C) tephra at about 35 ka. The end of the Seho period also varies between different workers but is commonly placed between 8 and 9 ka (Morrison, 1991; Benson et al, 1992).

Lake level fluctuations within the Seho cycle have also received a great deal of attention through the years, particularly the timing of the Seho highstand. However, at present several chronologies exist all of which are based on some number of the radiocarbon dates referred to above (e.g. Thompson et al, 1986; Morrison, 1991; Benson et al, 1995) and which place the timing of the highstand between about 12.5 to 14 ka.

An important point to keep in mind when discussing the lake level and depositional history (e.g. shoreline, deltaic and deep water deposits) of Lake Lahontan is that there is not a common history for all of the different subbasins in the system. Instead, each of the seven distinct subbasins have their own unique histories and shoreline records which are only shared with adjacent subbasins when lake level rose above the adjoining sill. Only when Lake Lahontan rose above its highest sill at Adrian Valley (~1311 m) did the entire Lake basin share a common shoreline record (Figure 1).

Controls on Shoreline Development

There are several factors or conditions which control the type of landforms, features and deposits found in coastal or shore settings. The factors have been delineated primarily from studies pertaining to ocean coasts, but we submit that these same controls are broadly applicable to large lacustrine systems such as Lake Lahontan. The main differences being that lakes experience negligible diurnal tidal fluctuations and generally have a restricted fetch. Regardless, there are several controlling factors in common which include 1) preexisting morphology of the shoreline and adjacent lake bed 2) sources, physical properties and rates of sediment supply 3) climatic and wave energy environment and 4) rate and direction of lake-level change (Forbes and Syvitski, 1994). Each of these factors may act alone or in concert to produce unique combinations of shore features and deposits.

The bedrock geology of the Jessup Embayment has not been mapped in detail, but generally consists of Tertiary volcanic and sedimentary rocks resting on a basement of Triassic and/or Jurassic metasedimentary and metavolcanic rocks which are intruded by rhyolite plugs and dikes (Wilden and Speed, 1974). The distribution of the rhyolite units is later used to delineate shore drift directions within the Embayment.

The preexisting morphology of the Jessup Embayment was that of a fairly broad alluvial valley narrowing towards its head and which slopes southeastward at about 2-3 degrees along its central axis. The Embayment extends about 3 km from its mouth to the head of the bay and is bordered on the southwest and northeast by prominent headlands or islands depending on lake level (Figure 2). The central part of the Embayment has abundant sand and gravel deposits of unknown depth but extending to at least 10 m in some locations. At the highstand the plan view of the shoreline was very irregular with many headlands and pockets formed by the superposition of a horizontal water plane on a landscape sculpted predominately by subaerial processes.

Most of the lacustrine sediment in the Jessup Embayment was probably derived from alluvial and/or lacustrine deposits existing within the Embayment prior to the Seho Lake cycle, produced by erosion of bedrock within the Embayment by wave action or introduced from areas upslope of the Embayment during the Seho Lake cycle. This latter process likely had a relatively minor input because deltaic deposits grading to the highstand or any other lower lake level appear to be absent in the Jessup Embayment.

Most of the surficial sediment in the Embayment has been separated into two mapping units referred to as the beach gravel (Qsg) and the beach and offshore sand (Qss) units, both of which were deposited during the Seho Lake cycle (Figure 2). There are no appreciable fine-grained (silt and clay), deep water lacustrine units exposed within the Embayment. However, the Jessup Embayment is fronted by extensive fine grained playa deposits of the Carson Sink (Qp₁). The distribution of the Qsg and Qss units appear to be controlled by both local slope and longshore transport. The Qss unit is mostly found within the central part of the Embayment where the slope is relatively gentle (~2°). The unit is also found below headlands and islands within the Embayment where the local slope shallows from relatively steep slopes (6-13°) to relatively gentle slopes (~2-4°). There are

also isolated patches of the Qss unit which are found on gently sloping terrace treads formed on overall steeper slopes such as the southwest side of the east island (Figure 2). The dominant clast size in the Qss unit is medium to coarse sand with some well rounded pebbles and angular fragments of branching tufa.

The Qss unit is interpreted to be the result of offshore movement of sand due to wave action on a steep coast. Roy et al (1994) report that simulation modeling suggests that sand moves offshore on submarine slopes steeper than about 1° and onshore for slopes much less than 1° . These trends operate despite the direction of relative sea-level movements. The mechanism which facilitates this offshore movement of sand is the suspension of sand size material by direct wave action (Roy et al, 1994). This process therefore represents a winnowing effect on sediment in the wave affected zone and the resultant concentration of pebbles and larger clasts in many of the shoreline deposits and landforms within the Embayment. The barriers formed of unit Qss were built during slight stillstands in the overall regression from the Seho highstand.

The beach gravel (Qsg) unit is found throughout the Embayment on steep as well as gentle slopes. All of the highstand depositional features are composed of Qsg. Below the highstand, the unit is arranged into barrier ridges as well as sheet-like bodies on steeper slopes. The surficial sediments of the terraces are also composed of the Qsg unit. Clast size ranges from sand through pebbles, cobbles and occasional boulders depending on the local energy environment and sediment availability. Sand is generally a minor component in the unit and was probably winnowed and moved offshore through direct wave action as discussed above.

The distribution of a third lacustrine unit (Qbs) is limited to a spit-like feature in the southwestern part of the mapping area (Figure 2). The feature is known as the boulder spit and consists of large blocks (up to 1.5 m) of basalt arranged in a recurved fashion around a bedrock core. Some of the blocks exhibit rounding and the interior of the deposit is cemented by dense laminated tufa and the exterior is coated with branching tufa (terminology from Benson, 1994). The north side of the deposit slopes at steeper than the angle of repose but is essentially held up by the tufa cement. There are two terraces on the upper part of the deposit which imply that at least the upper surface has been affected by waves. The top of the Qbs unit lies about 5 to 10 m below the crests of adjacent highstand barriers and appears to lie stratigraphically beneath the Qsg and Qss units which surround the boulder spit. We interpret the boulder spit to be a wave formed deposit that probably predates the Seho Lake cycle.

The present climate of the Lahontan basin varies according to both latitude and elevation but can generally be characterized as semiarid to arid with hot summers and cool, relatively wet winters (Houghton et al, 1975). During the Seho Lake cycle, the climate was effectively much wetter, with inflow into the Basin exceeding evaporation (Mifflin and Wheat, 1979; Benson and Thompson, 1987b). Most attempts to model the pluvial climate in the Great Basin have called on some combination of cooler temperatures and increased precipitation (Mifflin and Wheat, 1979; Benson and Thompson, 1987b; Hostetler and Benson, 1990). For an excellent review of paleoclimate modeling efforts in the Great Basin see Benson and Thompson (1987b).

Although reasonable estimates have been made about the temperature, precipitation, amount of cloud cover and other paleoclimate indicators during the Seho Lake cycle, little is known about wind conditions during this time. Therefore, the wave energy environment that existed in the Lahontan basin during the Seho Lake cycle can only be described in qualitative terms. Judging by the size, degree of development, and large clast sizes (up to

50 cm) within many of the shore features, it is clear that Lake Lahontan was a place where large waves were generated by strong winds.

Prior to the Seho Lake cycle, the hydrologic condition of the Lake Lahontan basin was probably much as it is today with relatively small lakes occupying the lowest parts of the Basin (Morrison, 1964, 1991). Therefore, the Seho Lake cycle can be viewed as a large scale transgression to the highstand and a similarly large scale regression back down to the Basin floor. There were certainly oscillations within these overall trends and the rates at which these major and minor fluctuations in water level occurred is not well known due to the relatively coarse resolution of existing lake level chronologies (Benson et al, 1995). However, general rates of transgression and regression can be estimated using Benson et al's (1995) lake level curve (Figure 3). The curve was developed for the western subbasins which were connected to the Carson Sink subbasin when lake level was above the sill at Fernley (~1265 m). Only when the lake was above the level of this sill did the two subbasins have a common history. With this constraint, the steep rise in the lake level curve between 1265 m and 1335 m during the time period from about 14.8 to 14.1 ka represents a lake level rise of approximately 100 mm/yr. Similarly, the drop from the highstand to the Fernley sill during the time period 13.6 to 13.2 ka represents a regressive rate of approximately 175 mm/yr (Figure 3). Both of these rates are extremely rapid, and imply that the shore features formed during both the transgressive and regressive stages were created rather rapidly, as the Lake probably did not stabilize at many of the shoreline levels for any length of time.

Shoreline Types

Wave cut terraces

The most common type of shoreline in the Lahontan basin is the wave-formed terrace. Terraces are characterized by gently lakeward sloping platforms bordered at their upper margin by an eroded cliff (Gilbert, 1885). The juncture between the platform and upper cliff is known as the shoreline angle and the line formed by this juncture is almost always horizontal. The horizontal nature of terraces is what makes pluvial shorelines so distinctive when viewed from a distance.

The word terrace refers to the morphology of the feature and not to the material in which it is formed. In the Lahontan basin, a spectrum exists where terraces are cut into bedrock as well as thick accumulations of surficial material. Even though terraces are usually considered erosional landforms, many terraces have a depositional component to them. Figure 4 shows a cross section through a cut-and-built terrace as defined by Russell (1885). Note that there is a wedge of debris built out and extending the width of the terrace tread. Only the eroded bedrock should be considered a wave-cut platform. Oftentimes, the built part of the terrace is better developed than the cut or erosional part. However, an exposure through the platform is usually required to discern the relative importance of each of the components. When a terrace is excavated in surficial material it is very difficult to discern the built part of the terrace and hence, this discussion applies mainly to terraces formed on slopes with a thin depositional cover.

Cut-and-built terraces are the most characteristic form of terrace in the Lahontan basin (Russell, 1885) and are usually found on bedrock headlands and islands. In the Jessup Embayment, cut-and-built terrace treads are commonly composed of cobbles and boulders up to 50 cm in diameter and are often cemented by both dense laminated tufa as well as branching tufa. Usually the branching tufa is found on the exterior of the terrace tread and lower riser with the dense laminated tufa forming a cementing matrix within the terrace tread cobbles and boulders. The boulders and cobbles concentrated on the terrace

tread probably form as a lag deposit where the finer material is either moved directly offshore or along shore by wave action. These types of terrace treads may be likened to the outer boulder frames of Bluck (1967) and Carter et al (1990a, 1990b).

The most prominent of the cut-and-built terraces in the Embayment is found about 5 to 7 m below the crests of the highest depositional barriers. This terrace is best developed in the southwest corner of the Embayment and on the west side of the north island where it is cut into bedrock, ranges up to 40 m in width, slopes gently lakeward and usually has a well-developed cliff at its landward edge (Figure 2). The cliff on the west side of the Embayment as well as the cliffs on the islands may have been excavated during the formation of this prominent terrace and subsequently accentuated at later, slightly higher lake levels. In several places, the terrace appears to be composed of several closely spaced, vertically overlapping horizons of cemented beach gravel to boulders. As with lower cut-and-built terraces the interior of the deposit is cemented by dense laminated tufa and the exterior is coated by a thin layer of branching tufa. This terrace is probably broadly correlative with Russell's (1885) Lithoid Terrace and here is referred to as such.

Russell (1885) interpreted the Lithoid Terrace to represent the highstand of the penultimate lake cycle and due to its degree of development suggested that the penultimate highstand lasted comparatively longer than the most recent highstand. We have not made any observations contradicting this interpretation, and at this point can only state that the Lithoid Terrace predates the Sehoo highstand because small Sehoo highstand barriers are observed to lie on top of the Lithoid Terrace. If the Lithoid Terrace does date from the penultimate lake cycle, it is possible that it is correlative with the boulder spit (Qbs), as both lie at about the same elevation and are cemented by dense laminated tufa.

From observations within the Jessup Embayment and elsewhere throughout the Lahontan basin, it appears that the Lithoid Terrace and lower cut-and-built terraces are best developed and preserved on shores that are composed of intermediate to basic volcanic bedrock. Wallace (1977), in discussing the longevity of fault scarps in different materials, postulates that fractured bedrock can maintain steep slopes for up to 1 Ma in the Great Basin. However, he does not distinguish between different types of bedrock, but our observations suggest that at least in terms of shoreline terrace development, intermediate to basic volcanic flow rocks display the best development and preservation. This may be because of the tendency of andesite and basalt to break into blocks and not disintegrate into small pieces. Because of their blocky weathering habit and relative resistance, there is probably a correspondingly low production rate of colluvium on these types of shores which would act to mute the morphology of cut-and-built shorelines. Cementing of the terrace treads by dense laminated tufa also plays an important role in the longevity of cut-and-built terraces.

It is possible that the development and preservation of terraces is accentuated by the repeated effects of multiple lake cycles. The longevity of terraces formed in intermediate to basic volcanics may be such that the amount of time it takes to completely mute these terraces by weathering, erosional and depositional processes is longer than the time between major lake cycles. Therefore, each successive lake cycle would tend to accentuate the form and development of the terraces. This concept is referred to as polycyclic shorelines and can also be extended to the rounding of beach gravel (polycyclic gravel) where the degree of rounding may be the result of more than one major lake cycle. The Lithoid Terrace may be an example of a polycyclic shoreline with its major development taking place during the penultimate or an earlier lake cycle and accentuation taking place during the Sehoo Lake cycle. However, the surficial sediments were mapped as Qsg because they were at least wave affected by the Sehoo Lake (Figure 2).

The lower terraces do not form exclusive horizons separate and unique from the barriers in the Embayment but instead trend into barriers in at least two locations. In the southwest part of the Embayment is a cusate barrier complex which has several erosional terraces on its southeastern flank (Figure 5) that trend directly into barriers at 1235 and 1227 m. As in all cases observed throughout the Lahontan basin, where erosional terraces trend into adjacent depositional features (spits, barriers, etc.), the crest of the depositional feature usually lies one to three meters above the shoreline angle formed by the terrace tread and adjacent upper riser or cliff. This same phenomena was observed in the Lake Bonneville basin by Gilbert (1890, p. 122-125) and attributed to the idea that the shoreline angle approximates still water level of the lake whereas the crests of constructional features represent the effects of storm deposition.

Constructional and beach barrier features

Depositional or constructional shorelines are also prevalent in the Lahontan basin and are commonly in the form of some sort of beach barrier. The essential distinction between coarse clastic barrier features and beaches is that all barrier types have identifiable crests and commonly have a well-defined backslope or back barrier depression (Carter and Orford, 1993). In contrast, beaches always slope lakeward. For the purposes of this paper, depositional shorelines are further subdivided into spits, pocket barriers, cusate barriers, loop barriers and other types according to the classification scheme outlined in figure 6. Although depositional shorelines can be broadly classified into drift or swash aligned features, most shorelines probably exhibit the effects of both processes at different stages of their development (Orford et al., 1995). Excellent examples of most of the above types of constructional landforms can be found in the Jessup Embayment. Prominent barrier ridges occur from an elevation of 1227 m up to the highstand at about 1340 m. Most of the larger and more continuous barrier ridges are located in the central part of the Embayment where the average slope is about 2 degrees (Figure 2).

The sequence of barriers from about 1328 m down to about 1227 m (Figures 2 and 5) are interpreted as recessional barriers and the surface on which they are formed as a recessional strand plain (Roy et al, 1994), even though the latter is on a smaller scale than its marine analog (Figures 2 and 5). Where the slope increases past about 5 degrees, as on the islands and headlands in the Embayment, it appears that cut-and-built terraces are the dominant shoreline type. This points to a slope control on the type of shoreline formed in the lacustrine environment. However, other factors are probably involved which dictate the type of shoreline formed including sediment character and availability (both locally derived and longshore drift derived) and irregularities in the shore that may act as sediment traps to longshore transport.

Sediment availability and the rate of lake level change (rising and falling) play important roles in the development and morphology of barrier ridges. For a stable or rising water level and limited sediment supply, swash aligned coarse clastic barriers tend to migrate landward (Carter and Orford, 1984; Orford et al, 1991b; Orford et al, 1995) through a process called barrier rollover. This is accomplished by overwash processes where material is moved from the lakeward side of the barrier, over the crest and deposited on the backside of the barrier (Carter and Orford, 1984; Orford et al, 1991a, 1991b). In contrast, overtopping is when the onrush of sediment and water just reaches the crest of the barrier, causing deposition at this location. Whereas overwashing leads to crestal breakdown and barrier migration, overtopping leads to crestal buildup and increased stability (Orford and Carter, 1982; Orford et al, 1991a, 1991b). Both of these processes lead to a net shoreward transfer of coarse clastic material. However, certain wave

environments and the effects of gravity can move some sediment lakeward down the front of the barrier (Bluck, 1967). Another reason for the landward migration of coarse clastic barriers is that breaking waves probably generate larger tractive forces directed landwards than the backwash does directed lakeward. The difference in the magnitude of tractive forces can in part be explained by the lesser volume of the backwash due to infiltration of water into the barrier (Carter and Orford, 1993). The net effect of this difference results in the movement of coarse sediment in a landward direction.

The Progradational Barrier Complex

A progradational barrier complex (PBC) is defined as a shore feature where multiple barrier ridges occur in close proximity to one another and at about the same elevation, which is indicative of a fairly stable water level and continued sediment supply (Orford et al, 1991a). A well developed PBC is located at the head of the Jessup Embayment about 6 to 8 m below the elevation of adjacent highstand features (Figures 5 and 7). In the complex, the individual ridges range in elevation from about 1332 m to about 1334 m and can be seen to truncate one another in plan view (Figure 7). The highest ridge is in the center of the complex with lower ridges both lakeward and landward. We interpret this relationship to indicate that water level was fairly static or fluctuated around some small elevation range during the building of the complex. The highest ridge in the complex (~1334; Lower Barrier 0 in figure 5) is not perfectly horizontal along its crest but instead has a noticeable trough about 50 cm deep and 15 m wide oriented perpendicular to the crest. We interpret this trough to be a paleo-overwash channel and call attention to this feature to make the point that the crests of both swash and drift aligned features are commonly horizontal, but oftentimes have irregularities and sometimes slope toward their distal ends. The surface age of the PBC is < 15 ka based on ^{36}Cl exposure dating of clasts sampled from the upper two meters of the complex (Figure 8) (Fred Phillips, written comm., 1995).

An intermittent stream occupying the main wash of the Embayment has eroded the landward side of the PBC and provides a clear exposure along strike of the complex (Figure 7). The site is marked as figure 8 in figure 5 and a log of the exposure is provided in figure 8. The exposure reveals a complicated history of multiple drift directions and erosive events in an overall aggrading pile of coarse clastic beach sediments. The lowest part of the exposure displays horizontally bedded and well sorted sand (Unit 1) overlain by several packages (Units 2, 3, and 4) of coarse gravel tabular beds dipping to the south (15 to 25°) which we interpret to be foresets indicating north to south spit building. At about the 55 m mark, the south dipping tabular foresets are overlain by north dipping tabular foresets (Unit 5), indicating a local reversal in net shore drift (Figure 8). Units 4 and 5 are in turn truncated by a coarse cobble gravel horizon (Unit 6) with a planar erosive base at about 1327.5 to 1328.5 m. Unit 6 fines upward into a well stratified pebble, sand and cobble zone (Unit 7) from about 1328.5 to 1329.5 m. On the log of the exposure (Figure 8), the apparent dip directions of Unit 7 are both to the north and to the south. However, these apparent dip directions are an artifact of the trend of the exposure. The true dip directions of this horizon and the horizons above are easterly and the apparent change in dip directions occurs because of the bend in section (Figure 8). The upper 20 to 40 cm of Unit 7 (~1330.5 m in the northern half of the exposure) is well cemented which we interpret to represent beachrock formed when this surface was the active beach. The surface of the beachrock is coated by a thin layer of branching tufa that coats only the upper parts of clasts embedded in the beachrock.

In the southern half of the exposure a coarse cobble to boulder horizon (Unit 8) cuts out the beachrock and tufa layer. Clasts in Unit 8 range up to 30 cm and are generally well-rounded. Within Unit 8 are clasts that have only one side coated with the branching tufa indicating that these clasts were probably eroded from the underlying beachrock and tufa horizon (top of Unit 7). Unit 8 is also present in the northern half of the exposure, but lies on top of and has not eroded the beachrock and tufa horizon (Figure 8). Above Unit 8 lie a series of pebble to cobble gravel horizons (Units 9 through 12) that are mostly clast supported but some have a sandy matrix filling the interstices between clasts. Both shape and size sorting are well developed in these upper horizons.

The surface ridges of the progradational barrier complex are composed of Units 8 through 12 which are exposed in the long section (Figure 8). The ridges and upper few meters of the progradational barrier complex were deposited during the regressive phase of the Seho lake cycle. This interpretation is supported by crosscutting relationships where a regressive spit (feature 2b, Figure 7) to the west of and at about the same elevation as the PBC (feature 3, Figure 7) truncates the distal end of a spit (feature 1, Figure 7) that began building during the highstand and was lengthened as lake level began to recede. The regressive spit in turn predates the barrier ridges of the complex because it is behind (landward of) the complex and so would be sheltered from strong wave action (Figure 7). Therefore, the barrier ridges also postdate the highstand. However, AMS radiocarbon dating suggests that the lower part (below about 1330 to 1331 m) of the PBC was deposited prior to the highstand. The contact between the transgressive (lower part of the exposure) and regressive (upper part of the exposure) deposits may be represented by the beachrock and branching tufa horizon.

Dating specific horizons within the PBC, combined with known elevations provide precise temporal and elevation estimates of lake level because the dated layers represent the shore of the Lake during the times of deposition. We collected tephra samples from three horizons within the lower part of the barrier complex exposure as well as gastropod shells from two of the same horizons. The tephra samples were taken from a horizontally bedded sand layer near the base of the exposure (Unit 1) (~1326 m), from a conspicuous fine grained layer (within Unit 2) sandwiched between the lower packages of tabular foresets (Units 2 and 4) (~1326 to 1326.5 m) and from just below the beachrock horizon about 15 m from the north end of the log (Unit 7) (~1330.5 m) (Figure 8). The three samples contained from 7 to 20 % volcanic glass shards so are not true ash layers, but rather ashy clastic sediments (Andrei Sarna-Wojcicki, written comm. 1995). The upper sample was collected from the matrix between coarse pebbles to cobbles and so has experienced some amount of reworking within a high energy beach environment. The middle and lower samples were probably also reworked because they too have low concentrations of glass shards.

The glass shards in all three of the ashy clastic sediment samples best correlate to one another and to a group of tephtras known as the Walker Lake-Negit Island Causeway set of "proto" Mono Craters layers which are estimated to be between ~65 to ~80 ka in age (Andrei Sarna-Wojcicki, written comm. 1995). These correlations indicate that the tephra layer or layers were originally erupted in early Wisconsin time and not during the late Wisconsin or Seho time. However, it is clear from the sedimentology of the layers as well as the glass shard concentrations that all three of the ashy clastic sediment horizons have been reworked and do not represent original airfall. The question is, how much time elapsed between the original eruption of these tephtras and their incorporation into the PBC?

Data bearing on this question are in the form of AMS radiocarbon dates from gastropod shells collected from the ashy clastic sediment layers. We collected *Vorticifex* (*Parapholix*) *solida* shells (Burch, 1989) from both the upper and middle ashy clastic layers for radiocarbon dating and additional shells from throughout the section for X-ray diffraction studies. The dates were provided by the Swiss Federal Institute of Technology in Zurich. Shells from the upper ashy clastic sediment layer date from $13,110 \pm 110$ yr BP, while shells from the middle ashy clastic sediment layer date from $13,280 \pm 110$ yr BP (Figure 7) (Irka Hajdas, written comm., 1994). Although these radiocarbon estimates are in stratigraphic order, they certainly do not agree with the age estimates provided by the tephra correlations.

This situation implies one of two possibilities. Either 1) the shells and glass shards were deposited with the beach gravel sometime between 60 and 85 ka and then at some later date the shells were recrystallized, thereby incorporating young carbon and providing anomalously young ages or 2) the beach gravel, shells and glass shards were all deposited about 13.1 to 13.2 ka, which implies that the glass shards were derived from a deposit in the area. To ascertain which of these scenarios is most likely correct, we used X-ray diffraction to examine the composition of shells from each of the ashy clastic sediment layers, shells from the upper part of the sequence which we interpret to post date the highstand (≤ 12.7 ka) and shells from the shore of modern Pyramid Lake which we interpret to represent recently living examples of *Vorticifex*. According to Bøggild (1930), freshwater pulmonate gastropods which includes *Vorticifex* are composed of aragonite when living. All of the shell samples that we examined were also composed of aragonite and none were composed of calcite implying that the shells from the ashy clastic sediment layers have not been recrystallized and that their radiocarbon ages represent the age of the lower or transgressive part of the progradational barrier complex.

Natural Variability in the Height of Depositional Shorelines

The highstand constructional shorelines in the Jessup Embayment are a mixture of both drift and swash aligned features including spits, pocket barriers, and loop barriers (Figures 2 and 5). These features are the clearest examples of transgressive depositional features within the Embayment simply because of their location at the top of the stack of shorelines. Figure 5 shows the results of detailed surveys of the elevations of constructional highstand features. The shorelines were measured using a Total Station surveying instrument which combines an electronic distance measuring (EDM) device with a theodolite. We used local benchmarks for elevation control. The accuracy of the instrument is within a centimeter on shots of up to a few kilometers, but because the benchmark elevations are reported to the nearest 0.1 foot, the precision of the measurements is assumed to be well within ± 0.1 m. Even though there is as much as 2.6 m of difference in the 10 highstand shoreline measurements (Figure 5), we submit that all of these features were built during the last highstand and the differences reflect natural variability in the height at which the crest of a depositional shoreline forms above a still-water plane. Natural variability in the height of shorelines is controlled by the size of waves reaching a particular shore which in turn is controlled by fetch, lake bottom configuration, geometry of the shoreline (e.g. embayment vs. headland) and the presence or absence of offshore obstructions (i.e., islands or shoals) (King, 1972). Shores that tend to have the highest elevated barriers, relative to still water level generally have areas of large (10's of kilometers), unobstructed fetch and moderately steep slopes approaching the shore.

The measured highstand shorelines can be further separated into swash or drift aligned features. Overall, the average height of the swash aligned features is 1339.8 m,

while the average height of the drift aligned features is 1339.0 m or about 0.8 m less. However, the lowest measurement was taken on a swash aligned pocket barrier (Shoreline F, figure 5) located to the north of the progradational barrier complex at the head of the Embayment. For reasons explained in the section above, we interpret that there was a smaller barrier complex at the same site as the progradational barrier complex prior to the highstand when Shoreline F was formed. The smaller barrier complex may have acted as an offshore obstruction and therefore dissipated wave energy at the location of Shoreline F during the highstand. The relatively poor development of Shoreline F also supports the hypothesis that wave energy was not as vigorous at this location as compared to the rest of the highstand shorelines.

Longshore Drift Directions and Their Determination

The net longshore drift direction near the head of the Embayment is clockwise or from southeast to northwest to northeast (Figure 2). This pattern was determined by several techniques including mapping of a distinctive lithology in beach sediments, interpretation of landforms, stratigraphy and tracing the size and sorting of clasts in the barrier deposits. A distinctive flow-banded rhyolite outcrops and is found in the sediments in the western part of the mapping area (Figure 2). The distribution of this rock type in the sediments was mapped near the head of the Embayment and is denoted by a coarse dotted line in figure 2. Several observations can be made about the distribution of the flow-banded rhyolite. First, the furthest east that it is found in alluvial sediments is the western part of the main drainage upstream from the progradational barrier complex (Figure 2). This rock type was not found in any of the alluvial sediments upstream from the high shorelines in the drainages to the east of the main drainage. However, the banded rhyolite can be traced in the recessional barrier ridges to the northwest side of the north island. Banded rhyolite is found in all of the high shorelines from Shoreline A through Shoreline E (Figure 5). To the north and east of these high shorelines the upper limit of distribution of the banded rhyolite descends to Recessional Barriers 3 and 4 at about 1320 and 1317 m, respectively. The high pocket barriers on the east side of the main wash (Shorelines G through J, Figure 5) do not have any banded rhyolite within them, but instead all of the sediment within each pocket barrier was most likely derived from local sources (Figure 5). The distribution of the banded rhyolite in both the high barriers and recessional barriers demonstrates that this distinctive lithology was progressively spread to the east in successively lower barriers during the recession from the highstand.

Another method used to interpret net shore drift directions was noting the directions that spits were built. Shorelines B and D (Figure 5) are both spits built during the highstand from SSE to NNW which is consistent with a clockwise shore drift pattern at the head of the Embayment. As discussed above, after the development of the regressive spit (feature 2b, Figure 7), the ridges of the PBC (feature 3, Figure 7) were emplaced, effectively sealing the head of the Embayment. The spit and crest of the PBC are at about the same elevation, indicating that at this location continuing sediment influx via longshore drift caused a change in the character of shorelines from being dominantly drift-aligned to being dominantly swash-aligned. However, the banded rhyolite was spread to the northeast (clockwise) in the PBC and lower barriers which is evidence that the apparently swash-aligned barriers also exhibit characteristics of drift-aligned features.

Lower Barrier 4 is generally composed of coarse gravel on the surface, but at a depth of about 130 cm changes to a well sorted sand. We interpret this lower sand as the beach and offshore sand unit (Qss) and further maintain that the upper gravel part of the barrier was drifted in from the west over the top of the Qss unit. Our interpretation that unit

Qss was formed by the offshore movement of sand when the lake was at relatively high levels is strengthened because the recessional gravel barrier stratigraphically overlies unit Qss.

The last piece of evidence we present for a clockwise drift direction at the head of the Embayment is the eastward fining of sediment in Lower Barriers 3 and 4. We have not performed particle size analyses of sediment along the crests of these barriers, but observations in the field indicate that the mean particle size decreases while the percentage of sand increases. All of the above observations and interpretations point to a dominant clockwise drift pattern at the head of the Embayment, although the morphology and orientation of some of the recessional barriers may also be used to suggest that they were dominantly swash-aligned features.

The clockwise drift direction in the Jessup Embayment and the fact that it faces south to southeast implies that wind and waves which built the highstand barriers and lower, regressive barriers were primarily coming from the south or southeast. This observation is at odds with those of Morrison (1964, 1991) who states that during Seho time, strong storm winds never came from the south or southeast. However, judging by the caliber of Seho beach sediments (up to 30 cm) and excellent development of shore features in the Jessup Embayment, strong storm waves did come from the south and southeast during Seho time.

Shoreline Processes at the Highstand

A small highstand pocket barrier with an enclosed playette behind it, herein referred to as the Jessup playette, is located in the northwestern part of the Embayment (Figure 2 and Shoreline C in figure 5). We excavated a 5 m deep trench perpendicular to the barrier and into the playette. The exposure enables a more detailed understanding of shoreline processes at the highstand and an estimate of its timing. Figure 9 shows a detailed topographic map of the trench site and locations of the trench and adjacent soil pit. The crest of the predominately swash-aligned barrier lies at an elevation of 1339.9 m and the surface of the playette is about 20 cm lower. The closed depression which was subsequently filled by the playette sediments was created by the emplacement of the barrier across this small wash. The extent of the drainage basin available to fill the closed depression is limited to the small hills surrounding the playette which have a combined area of about 6000 m² (Figure 9). It is possible that the relatively large drainage along the southwest part of the map (Figure 9) supplied sediment into the playette prior to the breaching of the barrier by the drainage, but does not seem likely.

The sediment deposited in the wash and the associated fan located to the south and east of the playette (Figure 2) consists of coarse cobbles to boulders of the banded rhyolite as well as intermediate to mafic volcanics. Even though this coarse sediment is post-pluvial in age, we find it unlikely that the caliber of the material moving down this drainage has changed very much since the highstand. The slope between the axis of the drainage and the top of the lagoonal sands near the bottom of the trench (measured in a straight line) is about 5°. This is approximately equal to the slope of the drainage adjacent to the playette. If the drainage spilled into the closed depression behind the barrier, there should be stratigraphic or sedimentologic evidence. However, there are no cobbles or boulders within the playette-fill sequence as described below. The coarsest material present in the playette-fill are rare clasts ranging in size to a few centimeters which appear to be sourced from the surrounding hillsides.

Figure 10 is a log of the trench exposure showing the relationships between the different packages of sediments used to interpret the history and timing of the highstand at

this location. We interpret the sediments in the exposure to record a single lake cycle at this elevation. The different packages of sediments will be discussed beginning with the oldest and ending with the youngest.

The oldest package of sediment exposed in the trench is located at the base of the exposure in the northwestern half of the trench (Figure 10). This massive, well-consolidated unit consists of poorly sorted, angular mafic volcanic clasts (≤ 10 cm) supported by a matrix of medium sand to gravel. Clasts tend to be somewhat concentrated at the unit's upper surface. Within the unit there are common fine to very fine root casts that are iron stained. Overall, the unit appears weathered but there is not an identifiable paleosol developed on its upper surface. This unit is interpreted to be alluvium which was deposited at some time prior to the Seho lake cycle and is probably relatively thin, although the base was not reached. The contact with the next overlying unit is sharp and slopes gently to the southeast.

The next younger unit is referred to as the lagoonal sands and consists of a wedge-shaped package thickening to the southeast to a maximum of about 50 cm (Figure 10). The base of the unit is poorly sorted, matrix supported cobbles and gravel grading upward to well-sorted fine to medium sand at the top of the unit. Gravel and cobbles within the unit are angular, hematite-stained volcanics and increase in abundance to the northwest. The upper sandy part has a greenish cast, possibly indicating that this unit is reduced. There is also limonite-hematite staining along common fine root casts and precipitated along horizontal bands. Cross-bedding (amplitude ~ 4 -5 cm) is present in the upper part of the unit with apparent local transport directions both to the southeast and northwest.

The lagoonal sands are interpreted to represent sediment deposited in a back-barrier lagoon during the highstand. The contemporaneity of the lagoonal sands and the highstand is demonstrated by the way in which the barrier gravels interfinger with the lagoonal sands (Figure 10). Cross-bedding within the lagoonal sands is interpreted to represent in part, the effects of overwash where a volume of sediment and water was washed over the crest of the barrier and into the lagoon generating local currents. Sedimentation continued in the back barrier lagoon after the time when the last waves washed gravel over the crest of the barrier and into the lagoon and is represented by the thickness of lagoonal sands (~ 12 cm) above the tail of barrier gravel (Figure 10). We interpret this relationship to mean that the lake did not recede immediately upon depositing the overwashed barrier gravels in the lagoon, but probably maintained a level above ~ 1336.5 m. A relatively high lake level and the high permeability of this type of coarse clastic barrier may have provided the mechanism by which water could seep from the lake into the lagoon, thereby maintaining a high water level in the lagoon (Carter and Orford, 1993).

The barrier gravels are contemporaneous with the lagoonal sands, as stated above, and tend to be well rounded, well stratified and well sorted within strata. There is also a tendency for the gravels to be sorted according to shape which is a common characteristic of beach deposits (Bluck, 1967; Carr, 1971; Orford and Carter, 1982; Orford et al, 1991a). Shape sorting is particularly evident in the southeastern part of the trench where there is a population of disc shaped clasts that range up to about 25 cm and are oriented parallel to the ground surface (dipping $\sim 11^\circ$ SE) (Figure 10). Although the barrier gravels are comprised of lithologies ranging from mafic to felsic volcanics and metasedimentary rocks, the lithology of the coarse disc population is limited to the banded rhyolite. Because the banded rhyolite tends to break into platy clasts, it was preferentially sorted to the exclusion of other more equidimensional shaped clasts due to hydrodynamic conditions within the surf zone (Orford and Carter, 1982; Orford et al, 1991a). The package of coarse disc shaped cobbles can be seen to truncate the finer gravel layers below them and, hence, probably represents

the last phase of barrier sedimentation at this site. In cross section, the barrier is arranged into somewhat tabular beds that dip lakeward ($8-10^\circ$) on the lakeward side of the barrier, flatten near the crest of the barrier and dip steeply ($33-34^\circ$) landward on the back of the barrier. This arrangement of dip directions can be likened to foresets, topsets and backsets according to the position within the barrier.

The steeply dipping ($33-34^\circ$) backsets are tabular in their central part and smoothly grade into a horizontal surface near their distal end in an asymptotic relationship (Figure 10). The tabular nature is in part defined by the alignment of platy clasts. Individual sand layers within the horizontal portion of the barrier tail can be seen to ramp up into the steeply dipping backsets. We interpret the sedimentologic relationships to demonstrate progressive accretion and migration of the backsets toward the northwest. Because the barrier gravel interfingers with the lagoonal sands, it appears that the entire barrier migrated to the NW probably through barrier rollover (Figure 10).

The hinge point where the topsets steepen into the backsets occurs at a position about 25 meters from the southeastern end of the trench (Figure 10). The steepness of the backsets ($33-34^\circ$), tabular arrangement and lack of erosional surfaces within and at the base of the package all suggest that these beds were deposited in a standing body of water and are analogous to Gilbert-type forsets in a deltaic environment (Gilbert, 1885). If water level in the lagoon was relatively low (< 1336 m) and the package of steeply dipping backsets was deposited on the dry, landward dipping part of the barrier, erosion, channeling and truncation of depositional surfaces would be expected within this package and at its basal contact with the lagoonal sands. However, no evidence of erosion or channeling was observed, but instead the beds are regular and relatively continuous thereby supporting the hypothesis that these sediments were deposited in a standing body of water. The water level in the back barrier lagoon during deposition of the gravels can be approximated by the elevation of the crest of the hinge point which is at about 1338.8 m (Figure 10).

Water may have entered the lagoon by several ways including direct precipitation, runoff from the surrounding hills, waves washing over the crest of the barrier or by water infiltrating through the barrier from the Lake. Because the drainage basin for the lagoon is quite small (~ 6000 m²), direct precipitation and runoff from the surrounding hills was probably small compared to the amount of input of water from the Lake. Direct overwash of waves probably contributed significantly to raising water levels in the lagoon during storm periods.

The volume of water moving through the barrier depends on the hydraulic conductivity of the gravel and the hydraulic head defined by the difference in water height between the lake and lagoon (Carter et al, 1989). Observations of the barrier gravel in the trench down to an elevation of about 1338 m indicate that the hydraulic conductivity is probably quite high due to the clast supported nature of the gravel and the relative lack of fine grained matrix. During storm surges, the absolute lake level at the shore is increased which would cause a concomitant increase in the hydraulic head and a possible increase in the water level of the lagoon, depending on the rate of infiltration through the barrier gravels. The rise in water level at the shore produced by the storm surge also increases the probability of overtopping and overwashing by storm waves which would also lead to increased water levels in the lagoon.

Based on the interpretation that water level in the lagoon was at a minimum of about 1338.5 m and the majority of this water was due to percolation through the barrier gravels from the lake, we estimate that the still water level at the highstand of Lake Lahontan at this location was somewhere between 1338 and 1339 m. This estimate

suggests that this and other highstand barriers in the Embayment rose from 0.2 to 2.8 m above still water level (Figure 5).

The youngest package of sediments exposed in the trench were deposited subsequent to lowering of the Lake. We refer to the sediments as playette-fill deposits and they consist of alternating layers of sandy alluvial wash or fluvial deposits separated by layers of fine sand, silt and clay (Figure 10). The alluvial layers tend to be less continuous across the exposure than the fine grained layers and consist of fine to coarse sand with rare concentrations of granules to several centimeters. The beds are generally tabular but several wedge-shaped beds which pinch out to the southeast are observed in the northwestern half of the playette exposure. Many of the sandy layers have erosive bases and common crossbedding. The amplitude of cross-bedding ranges from 3 to 7 cm and, in most instances, local transport direction is from NW to SE. Less common cross bedding indicating SE to NW transport is also present. The thickness of the beds range from several cm to about 10 cm. In some cases their upper contact with the finer grained layers is abrupt but in others the contact is gradational indicating a fining upward sequence. We interpret the sandy layers to represent alluvial or fluvial wash layers generated by local rainfall events and sourced from the surrounding hillsides.

Finer grained layers interbedded with the alluvial layers are generally massive with no apparent sedimentary structures. The particle size distributions of two of the more prominent fine grained layers were examined and found to consist primarily of silt, fine sand and clay in order of decreasing abundance (Table 1). The thickness of the fine layers range from a few centimeters up to about 8 cm for two of the thickest layers near the southeastern part of the playette exposure (Figure 10). Almost all of the fine layers have common, fine vesicular pores and vertical cracks. Many of the fine layers also have common fine to coarse vertical root casts. Although there is evidence for subaerial exposure and drying, there is no evidence for soil horizonation below ~ 1339 m other than the pores, cracks and root casts.

The fine grained layers tend to be more continuous than the coarse, sandy alluvial layers and many can be traced across the entire playette exposure (Figure 10). Most of the fine layers are relatively horizontal, but several dip gently to the SE and have as much as 20 to 30 cm of relief. The layers also tend to be less continuous near the top of the exposure. Based on the particle size distribution (Table 1), massive nature, vesicular pores, and

Table 1. Particle size distributions for two of the more prominent fine grained layers in the playette-fill sediments.

| Sample # | Sand (wt %) | Silt (wt %) | Clay (wt %) |
|----------|-------------|-------------|-------------|
| Jesp95-7 | 13.03 | 79.72 | 7.24 |
| Jesp95-3 | 15.10 | 74.00 | 10.90 |

vertical cracks, we interpret the fine grained layers to represent eolian dust deposits that were either deposited directly on the playette surface as it aggraded or settled out of columns of muddy water during ephemeral flooding of the playette surface. This latter explanation would provide a mechanism for concentrating the dust that had fallen on the surrounding hills above the playette (Young and Evans, 1986).

At the top of the playette-fill sediments, above about 1339 m is a zone of soil development delineated on the log (Figure 10) and characterized by strong platy structure with intersecting vertical cracks causing the soil to break into strong prisms. The soil is

dominated by silt sized material with lesser amounts of both fine sand and clay. The interbedded, sandy alluvial units present lower down in the stack do not appear to be present in the soil but may be masked by bio- or pedoturbation. The soil is also characterized by strong effervescence, indicating an accumulation of carbonate most likely derived from atmospheric sources (Reheis et al, 1989, 1995; Chadwick and Davis, 1990). The zone of carbonate accumulation appears to be roughly coincident with the zone of soil structure. In contrast to the upper part of the fill, all of the playette-fill sediments below about 1339 m do not effervesce, indicating that there is no accumulated carbonate in these sediments. We do not think that carbonate was leached from these layers after deposition because evidence is lacking of leaching and reprecipitation lower down in the stack.

Papke (1976) reported that CaCO_3 is common in Nevada playas and Chadwick and Davis (1990) proposed that the introduction of carbonate-bearing eolian material derived from wind erosion of playas such as the Carson Sink was a major factor in the formation of soils in the Lahontan basin. The presence of CaCO_3 in the upper part of the trench exposure coincident with the depth of soil formation fits the model of Chadwick and Davis (1990) where much of the soil formation was the result of the introduction of carbonate rich dust into the surface sediments of the playette. However, if we are correct in our interpretation that the fine grained layers in the lower playette-fill package also represent dust accumulation, it is perplexing that these dust layers do not contain CaCO_3 . Therefore, either the carbonate has been removed, which we find unlikely as explained above, or there was not carbonate in the dust when it was originally deposited. Because the age of the playette-fill sediments is known to be younger than 12.7 ka, the deposition of the dust layers within the playette may record dust deposition through the Holocene.

The lack of carbonate in the lower layers of dust implies that the source of dust may have changed through the Holocene. If this is true, during most of the filling of the playette, the dust that was present in the area and actively aggrading on the surface of the playette probably did not come from the Carson Sink playa. Instead, the dust may have traveled from further removed sources, but its origin is uncertain.

Timing and Magnitude of the Seho Highstand

A new minimum age constraint on the timing of the Seho highstand is provided by an AMS radiocarbon age estimate of $12,690 \pm 60$ yr BP on a camel bone found near the bottom of the trench (Figure 10). The distal end of a radioulna (fused front fore limb) and a metacarpal (foot bone) from a *Camelops hesternus* were found at the contact between the lagoonal sands and the playette-fill sediments (Figure 11). Amy Dansie from the Nevada State Museum identified the bones (Personal comm., 1995) and the age estimate was provided by Thomas Stafford from the University of Colorado, Boulder (Written comm., 1996). The metacarpal was found in front of and adjacent to the radioulna suggesting that these two bones were connected by soft tissue when deposited and buried. We interpret that the bones were deposited shortly after the death of the animal, probably within a matter of months to years, not decades to millennia. Hence, the bones were deposited and buried immediately upon the recession from the highstand, coincident with the change in sedimentation rate and style.

The bones were found at the contact between the lagoonal sands and the playette-fill sediments which we interpret to represent a major change in depositional processes and sediment sources (Figure 11). Whereas the lagoonal sands represent subaqueous deposition in a standing body of water, the playette-fill sediments represent periods of fluvial and/or alluvial deposition separated by periods of stability and dust accretion either by direct subaerial deposition or settling out of shallow, ephemeral water bodies. The

contact between the lagoonal sands and the playette-fill sediments is abrupt and marks the time of recession of Lake Lahontan from the highstand. There is no evidence of weathering or soil formation at this contact and so we interpret that the Seho highstand occurred immediately prior to $12,690 \pm 60$ yr BP, the age of the camel bones.

When combining the age of the camel bone with the two AMS radiocarbon dates in the PBC, we may place new constraints on the timing and magnitude of the Seho highstand. An advantage of these dates is that they all pinpoint a shoreline elevation for a specific time or range of time. The two shell dates ($13,280 \pm 110$ and $13,110 \pm 110$ yrs BP) were taken from beach deposits and therefore place elevational as well as temporal constraints on the location of the lake shore leading up to the highstand (Figure 8). The minimum limiting date ($12,690 \pm 60$ yr BP) on the camel bone combined with the elevation of water at the highstand (1338 to 1339) provides a minimum constraint on the age and elevation of the highstand in the Jessup Embayment (Figure 12). A maximum time constraint on the highstand is provided by the age of the upper shell date (13,110 yr BP) (Figures 8 and 12) because the deposits from which the shells were sampled from predate the highstand. Therefore, the timing of the last Lake Lahontan highstand was between about 13.1 and 12.7 ka, but was probably closer to 12.7 ka because we interpret this to be a closely limiting age constraint. The duration of the highstand is unknown but was probably on the order of decades to maybe one hundred years. This interpretation is based on the relative development of constructional barrier features around the Basin and the relative lack of erosional terraces formed in bedrock near the elevation of the highstand. This observation implies that the highstand was of long enough duration to form well developed spits and barriers but not long enough to form a well developed terrace.

Comparison of our data with Benson et al's (1995) most recent lake level curve implies the timing of the highstand was younger and the elevation higher than previously proposed (Figure 12). The differences between the two curves may in part be explained by the different materials used to estimate both the timing and magnitude of the highstand. Benson et al (1995) primarily used radiocarbon dates from inorganic carbonate (tufa) to constrain both the timing and magnitude of the highstand. However, the water depth at which tufa forms is not known so the upper limit of tufa growth is a minimum estimate of lake level. From field observations of over two hundred high shoreline localities throughout the Basin (Adams and Wesnousky, 1994, 1995), we have never observed tufa present at the high shoreline level. Commonly, tufa is found about 5 to 7 m lower. The lack of tufa observed on high shorelines may reflect our bias of examining constructional highstand features. Tufa is most commonly found on stable substrates in places that received high wave energy. However, we have not observed tufa on steep bedrock slopes or cliffs adjacent to and at the same elevation as highstand constructional shorelines. These observations may suggest that approaching the time of the highstand, lake level rise was so abrupt and the duration of the highstand so short that tufa did not have time to precipitate. Alternatively, the chemistry of the lake water may have changed enough during the brief rise to the highstand that geochemical conditions were not conducive to tufa precipitation. Both of these possibilities imply that Lake Lahontan received a sudden influx of water that caused a steep rise in lake level for a relatively brief amount of time, which is reflected in both Benson et al's (1995) curve as well as the data from this study.

Regression from the Sehoo Highstand

The physical record of regression from the Sehoo highstand is well displayed in the Jessup Embayment. There is a series of no less than 28 distinct barrier ridges formed as lake level dropped back down to the floor of the Carson Sink (Figures 2 and 5). The elevations of 12 of the more prominent ridges are shown in figure 5. We earlier used the lake level curve of Benson et al (1995) to estimate an overall regression rate of 175 mm/yr (Figure 3). By using the ages and elevations of packrat middens from the Winnemucca dry lake (WDL) subbasin (Thompson et al, 1986), the age of the highstand (~12.7 ka) from this study and assuming that the Carson Sink and WDL subbasins had similar dessication histories below the level of the Fernley sill, we can estimate that the lake fell from about 1338 m to about 1230 m in 630 years which corresponds to a rate of about 170 mm/yr. This estimate is essentially identical to the one using the curve from Benson et al (1995). Within the overall regression there were several elevations where the lake paused for a long enough period of time to build several ridges at about the same elevation. The highest of these is the PBC which is discussed above and lies about 6 to 8 m below adjacent highstand features. The next lowest group of ridges occurs at an elevation of about 1235 m and is located between lower barriers 10 and 11 in figure 5. A distinctive characteristic of these ridges is that they are completely covered with branching tufa and have abundant tufa heads and tufa towers (≤ 3 m) growing from them. The tufa covered barriers are designated Qss/t on figure 2.

The tufa appears to grow from a particular horizon as viewed in stream cuts through the barriers. Below the horizon, the beach gravel and sand is well cemented but there is little tufa present. Tufa heads are rooted in the horizon and clasts there are coated with branching tufa. The tufa is generally associated with the Qss unit, but in places also coats the Qsg unit. The surface width of the tufa coated barriers is about 250 m in the central part of the Embayment. However, near the headlands to the NE and SW (Figure 2), the tufa dies out only to reappear at the same elevation in other embayments to the north and south of Jessup. The common characteristics of these tufa bearing zones is their elevations and their locations near the mouths of broad alluvial valleys. We interpret the tufa-bearing ridges to represent a temporary stillstand in the overall regression where ground water was moving down the broad alluvial valleys and forced to come to the surface near the shore-lake interface. As the ground water moved through the lacustrine deposits in the upper parts of the embayments, it may have dissolved and moved calcium and/or carbonate in solution which was then precipitated when the ground water came in contact with the lake water. The reason that tufa did not precipitate downslope from rocky headlands is that there was not sufficient water moving through the thin sediments as there was in the broad alluvial filled valleys. The absolute age of the tufa coated barriers is not known.

After the formation of the tufa coated barriers the lake further regressed to some unknown elevation, but then rose again almost to the same level. Lower barrier 11 is at an elevation of 1234.6 m and stratigraphically overlies the tufa coated barriers (Figures 2 and 5). Whereas the surface of the tufa-coated barriers is completely covered with fragments of branching tufa, the surface and interior of lower barrier 11 is clean, well-washed beach gravel with only occasional fragments of tufa. After the lake built lower barrier 11, it regressed to about 1227 m where it built 3 barriers at about the same elevation (Figure 5). The highest of these is designated lower barrier 12. An interesting feature to note is that downslope from the lowest barriers is a sand sheet designated as Qss in figure 2. The deposit probably represents the offshore movement of sand during this minor stillstand, similar in respects to the thick deposits of sand in the upper, central part of the

Embayment. There are no lower barriers recognized between about 1227 m down to floor of the Carson Sink at about 1185 m, but this zone is dominantly covered by alluvium. Work is currently in progress to define the ages of the lower barriers.

Summary

This paper has delineated the lacustrine history of the Jessup Embayment of Lake Lahontan using a combination of geomorphic, sedimentologic, stratigraphic and age dating techniques. Modern process studies of coarse clastic barrier systems were also used to provide a framework within which the paleo-shorelines of the Embayment could be interpreted. The results of this study have confirmed and further elucidated the operation of similar shore processes in the ancient lacustrine record.

Other significant results of this study are 1) documentation of probable pre-Sehoo shorelines located about 7 to 10 m below the Sehoo highstand, 2) characterization and distribution of constructional shorelines and terraces in relation to sediment availability and local slope, 3) a classification scheme for constructional shorelines in pluvial lake basins is introduced, 4) the natural variability at which the crests of constructional shorelines form above a water plane was determined to be about 3 m and depends on fetch, lake bottom configuration, geometry of the shoreline and the presence or absence of offshore obstructions, 5) the net longshore drift pattern in the Embayment was clockwise indicating that large waves came from the south and southeast, 6) during the transgression to the highstand, lake level was at about 1326.5 m by 13.2 ka and at about 1330.5 m by 13.1 ka, 7) highstand constructional features in the Embayment average about 1339 m and date from the Sehoo highstand which reached an elevation between 1338 and 1339 m, 8) an AMS radiocarbon date on a camel bone from a lagoon enclosed behind a high pocket barrier indicates that the Sehoo highstand occurred at about 12.7 ka, 9) comparisons with the lake level curve of Benson et al (1995) indicate that both the magnitude of the highstand was higher and the timing younger than previous estimates, 10) discrepancies in the curves may be explained by the observation that tufa is not observed on highstand constructional features, but is found some meters lower, 11) the presence of carbonate-bearing dust at the surface of the playette enclosed behind the high pocket barrier and lack of carbonate dust at depth in the playette-fill sediments may indicate that dust sources have changed through the Holocene, 12) during the overall regression from the highstand, the lake formed at least 28 distinct barriers in the Jessup Embayment, 13) lake level paused long enough to form multiple barrier ridges at three different elevations which are at about 1332, 1235 and 1227 m, respectively, 14) tufa-coated barriers at about 1235 m probably reflect ground water interactions with lake water during a pause in the regression, and 15) after the formation of the tufa coated barriers, the lake receded but retransgressed to about 1234.6 m and then regressed to the floor of the Carson Sink.

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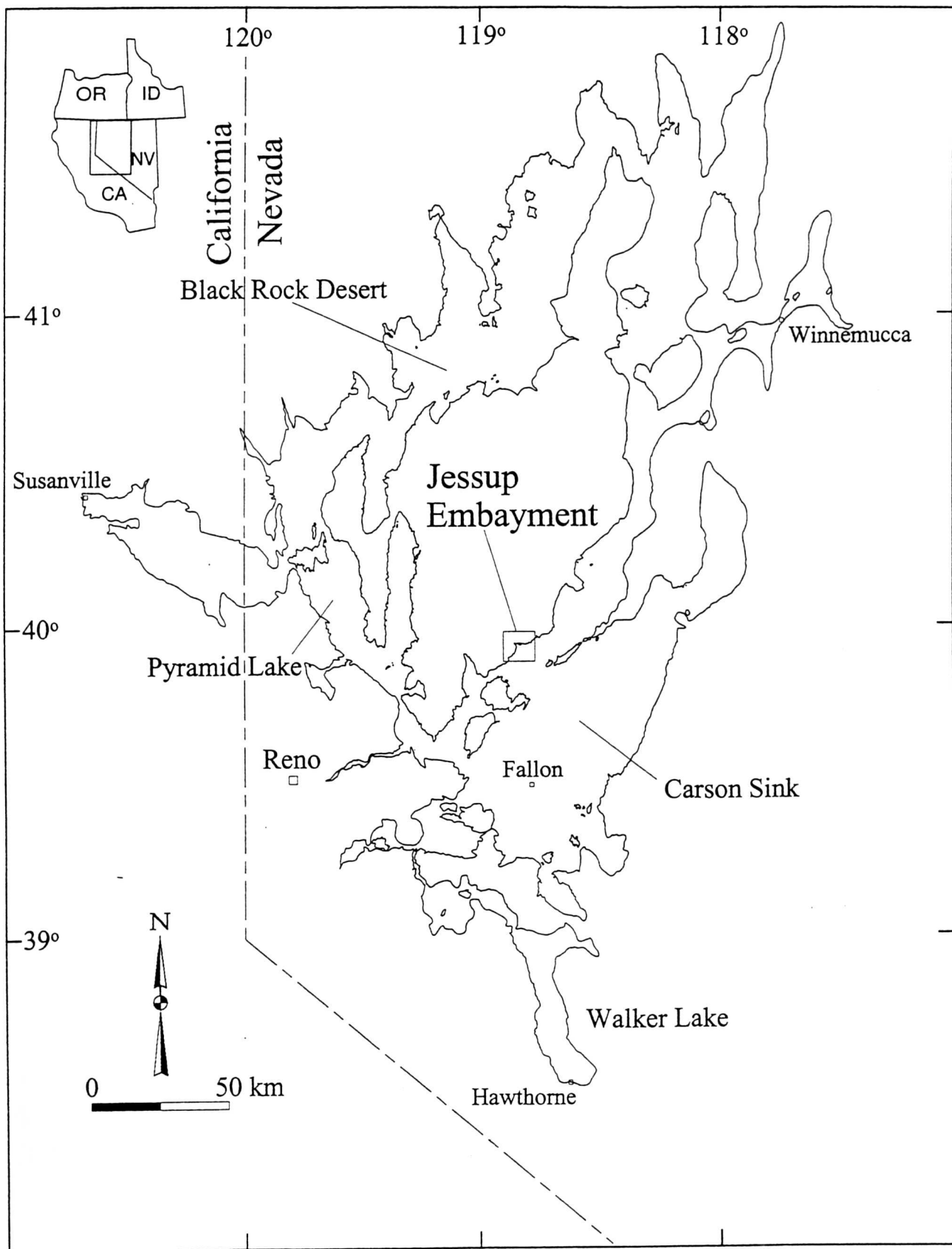
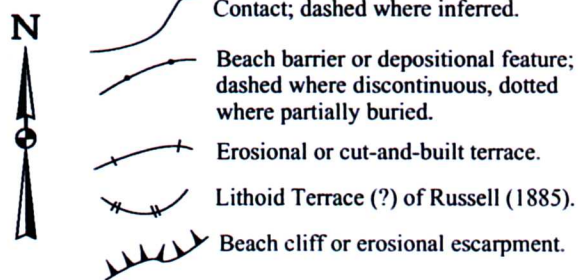
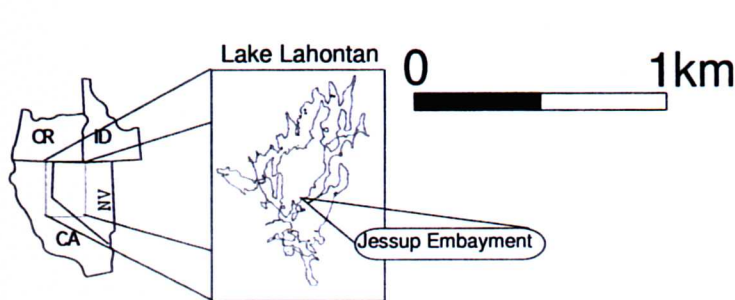
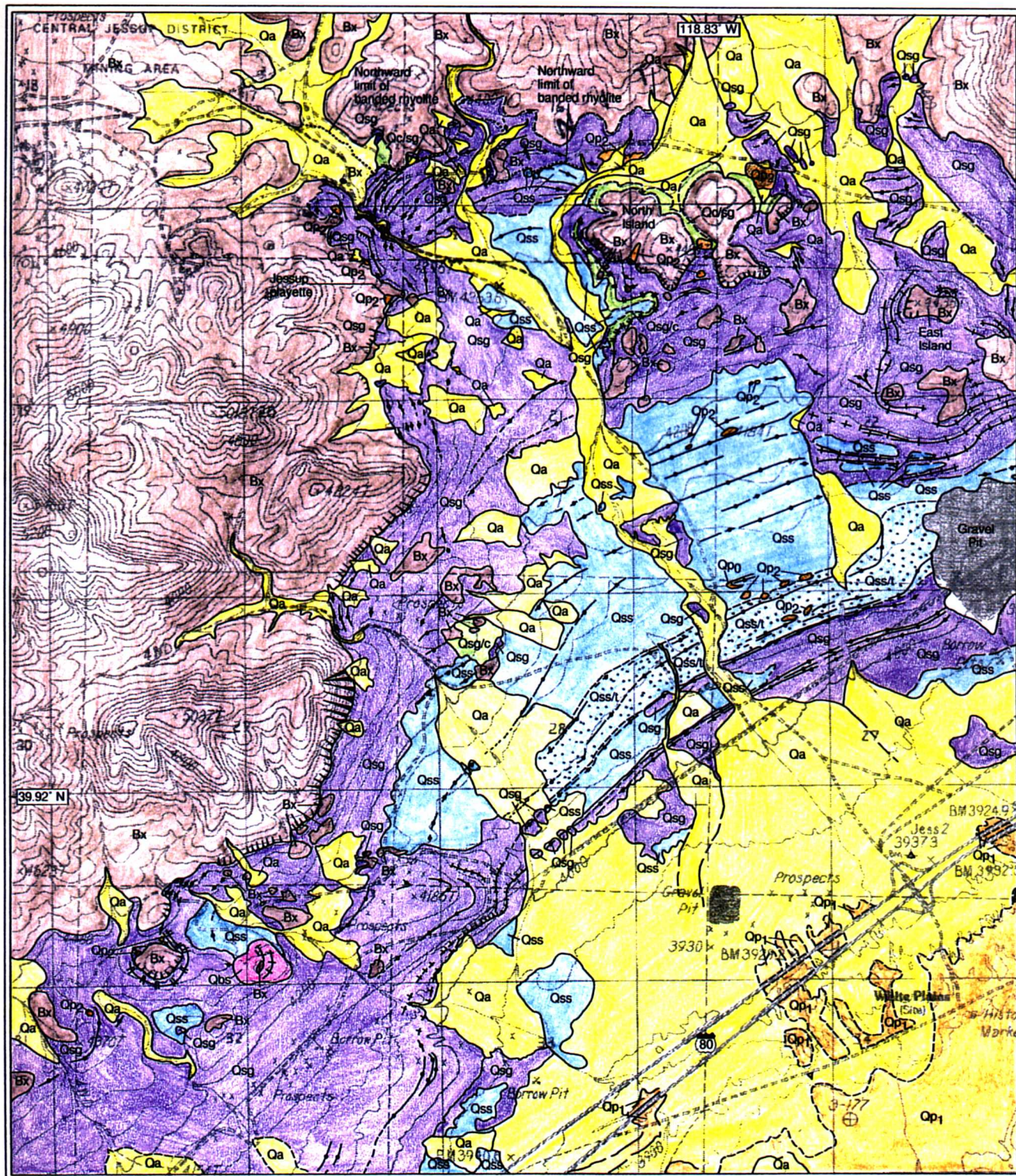


Figure 1. Map of Lake Lahontan as it appeared during the Seho highstand. The locations of the Jessup Embayment and other geographic features are also shown.

Figure 2. Geologic and geomorphic map of the Jessup Embayment showing the distribution of deposits and landform elements within the Embayment. Qa: alluvium of modern washes and fans. Qp₂: playette deposits. Qp₁: playa deposits of the Carson Sink. Qp₀: buried playette deposits predating Seho highstand. Qsg: beach gravel of Seho Lake cycle. Qss: Beach and offshore sand of Seho Lake cycle (may contain areas of Qsg). Qss/t: Surficial coating of branching tufa and occurrence of small tufa domes in unit Qss. May in places be associated with unit Qsg. Qsg/c: beach gravel of Seho Lake cycle mixed with lesser amounts of colluvium. Qc/sg: colluvium mixed with lesser amounts of beach gravel. Qbs: boulder spit of pre-Seho (?) age. Bx: Triassic and/or Jurassic metasedimentary rocks intruded by Tertiary rhyolitic plugs and dikes.

The northward limit of banded rhyolite in surficial sediments is denoted by a coarse dotted line near the head of the Embayment.



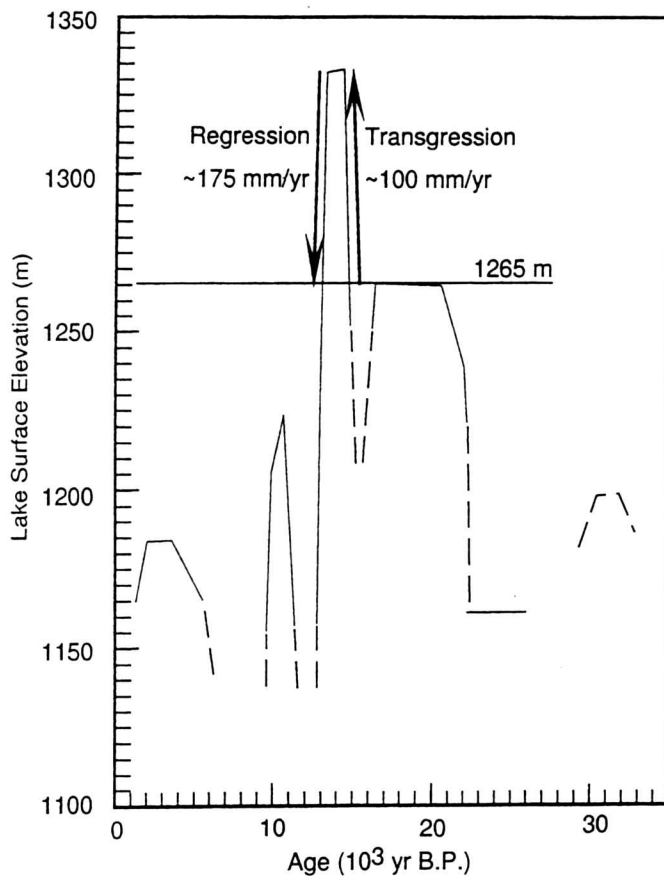


Figure 3. Lake level curve of Benson et al (1995) showing the estimated rates of lake level rise and fall above 1265 m when the Pyramid Lake and Carson Sink subbasins were connected.

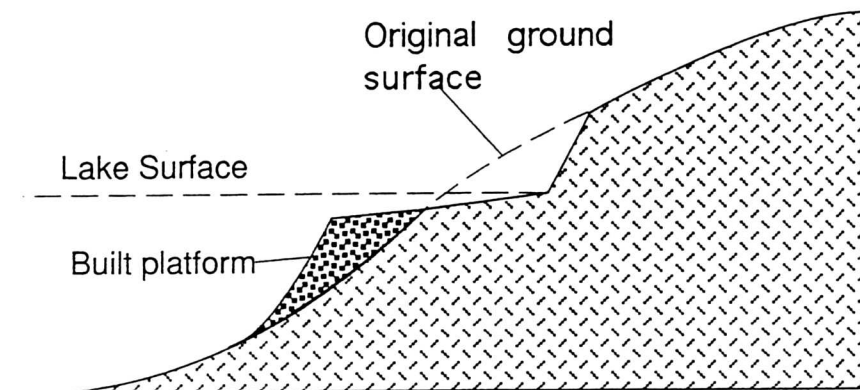


Figure 4. Idealized cross section through a cut-and-built terrace showing the relationship between erosional and depositional elements in this type of feature. After Russell (1885).

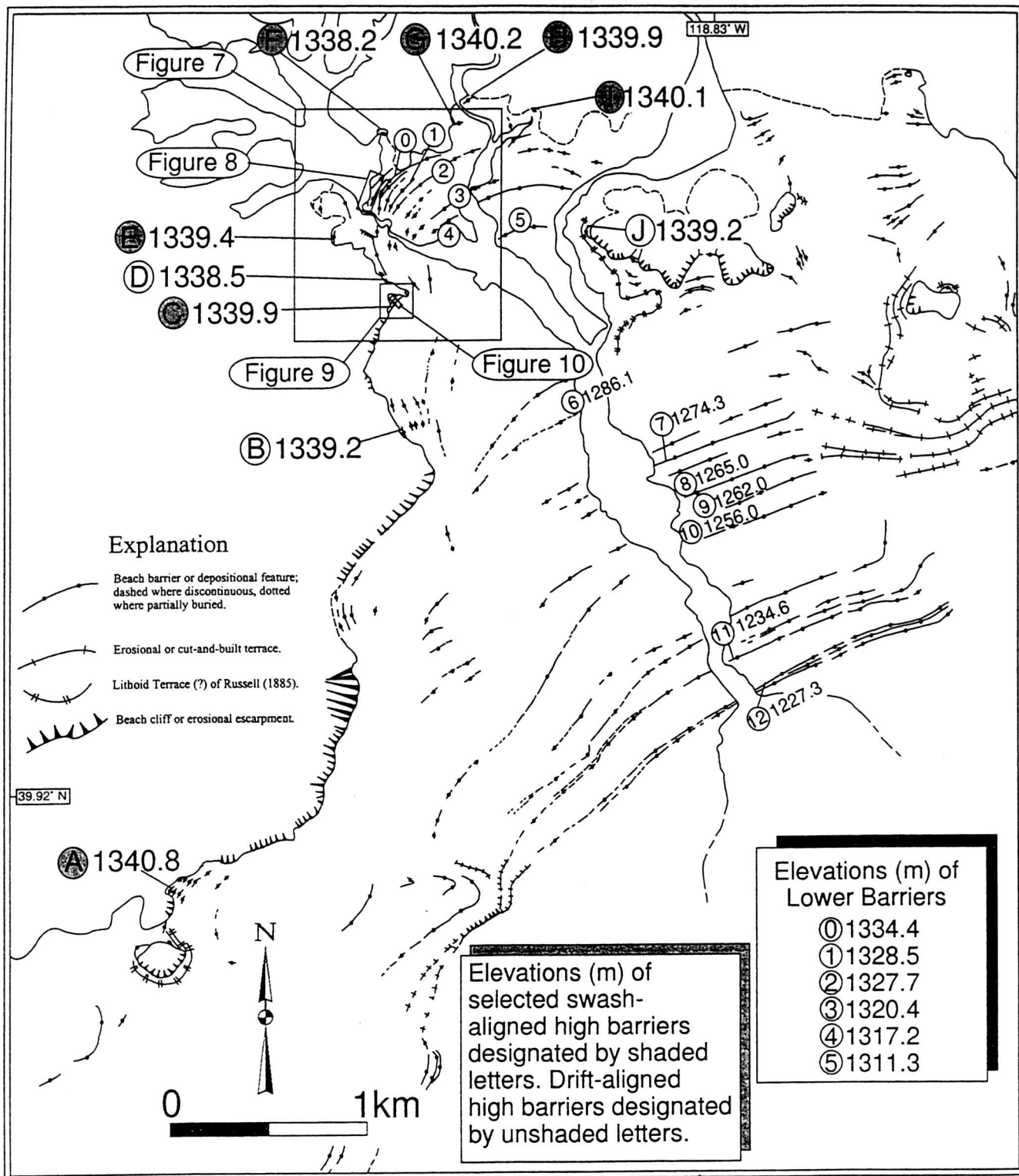


Figure 5. Location map showing shoreline elevations and figure locations. Swash-aligned highstand barriers and elevations are designated by shaded letters and drift-aligned highstand barriers and elevations are designated by unshaded letters. Recessional barriers are designated by numbers. We interpret all of the high shorelines to date from the Sehoo highstand. The difference in elevations (~2.6 m) is attributed to natural variability in the height of formation above a given water plane.

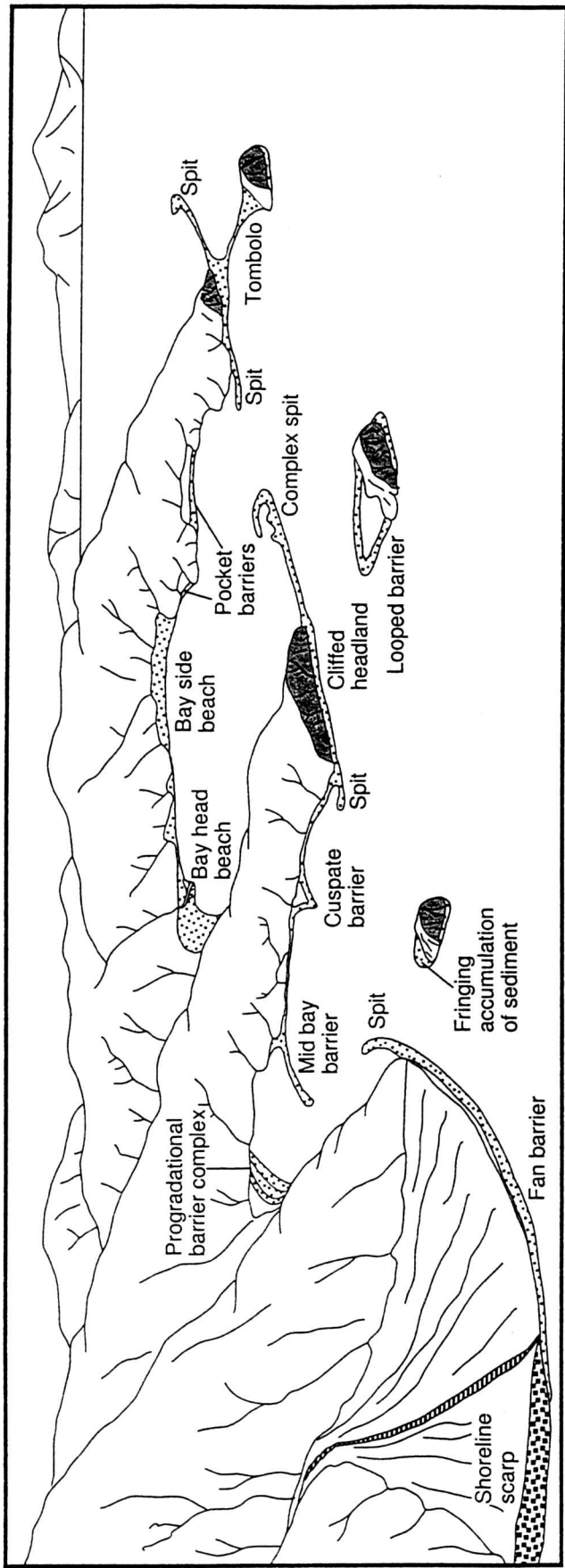


Figure 6. Shoreline classification scheme for pluvial features found in the Lahontan basin. After Strahler and Strahler (1992), King (1972) and Duffy et al (1989).

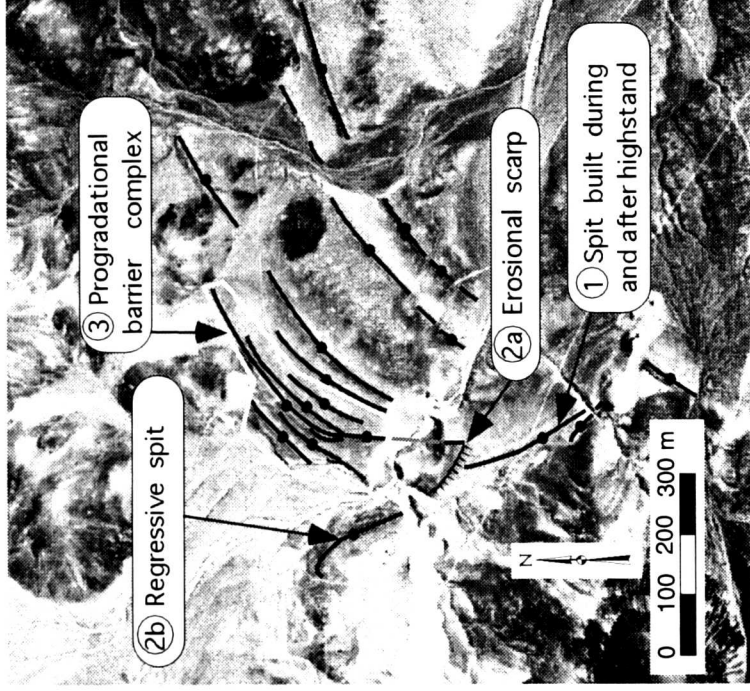


Figure 7. Aerial photo interpretation of barrier features at the head of Jessup Embayment (see figure 5 for location) showing cross-cutting relationships demonstrating surface ridges of progradational barrier complex (PBC) postdate the Sehoo highstand. Symbols are as in figures 2 and 5.

Feature 1 is a spit that began building during the highstand and continued to elongate to the north as lake level began to recede. Feature 2a is an erosional scarp that truncates the highstand spit and is probably contemporaneous with Feature 2b, which is a spit built from south to north at a level about 8 meters lower than adjacent highstand features. Feature 3 comprises the surface ridges of the PBC which also lie about 6 to 8 meters below adjacent highstand features.

We interpret these shorelines to indicate that first the highstand spit was built, and second, the end of the highstand spit was truncated and the recessional spit built. The ridges of the PBC were then emplaced across the head of the Embayment through continuing longshore drift and a relatively stable lake level. The lower recessional barriers were then formed as lake level continued its overall decline.

Long Section of Progradational Barrier Complex at the Head of the Jessup Embayment

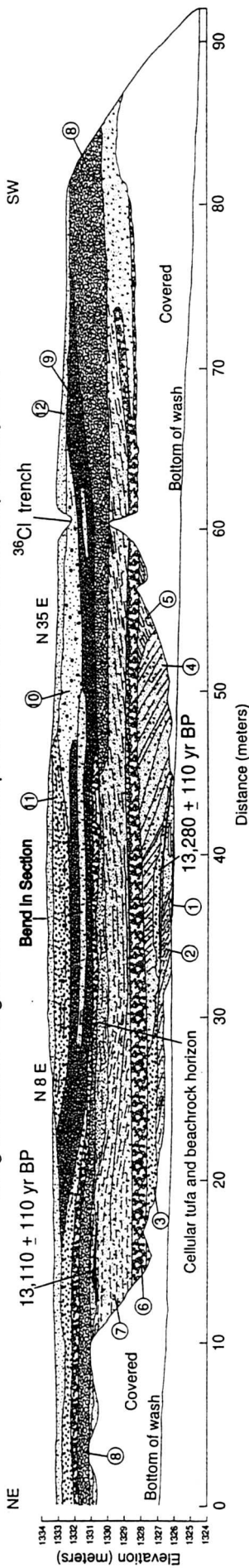


Figure 8. The progradational barrier complex (PBC) at the head of the Jessup Embayment showing main structural and sedimentary features exposed in a stream cut. AMS radiocarbon dates ($13,280 \pm 110$ yr BP and $13,110 \pm 110$ yr BP) on gastropod shells from the lower part of the section indicate that the package below the tufa and beachrock horizon predates the highstand at about 12.7 ka. Cross-cutting relationships of the surface barrier ridges (Figure 7) are interpreted to indicate that the sediment above the beachrock and tufa horizon post-dates the highstand. The numbered horizons are discussed in the text.

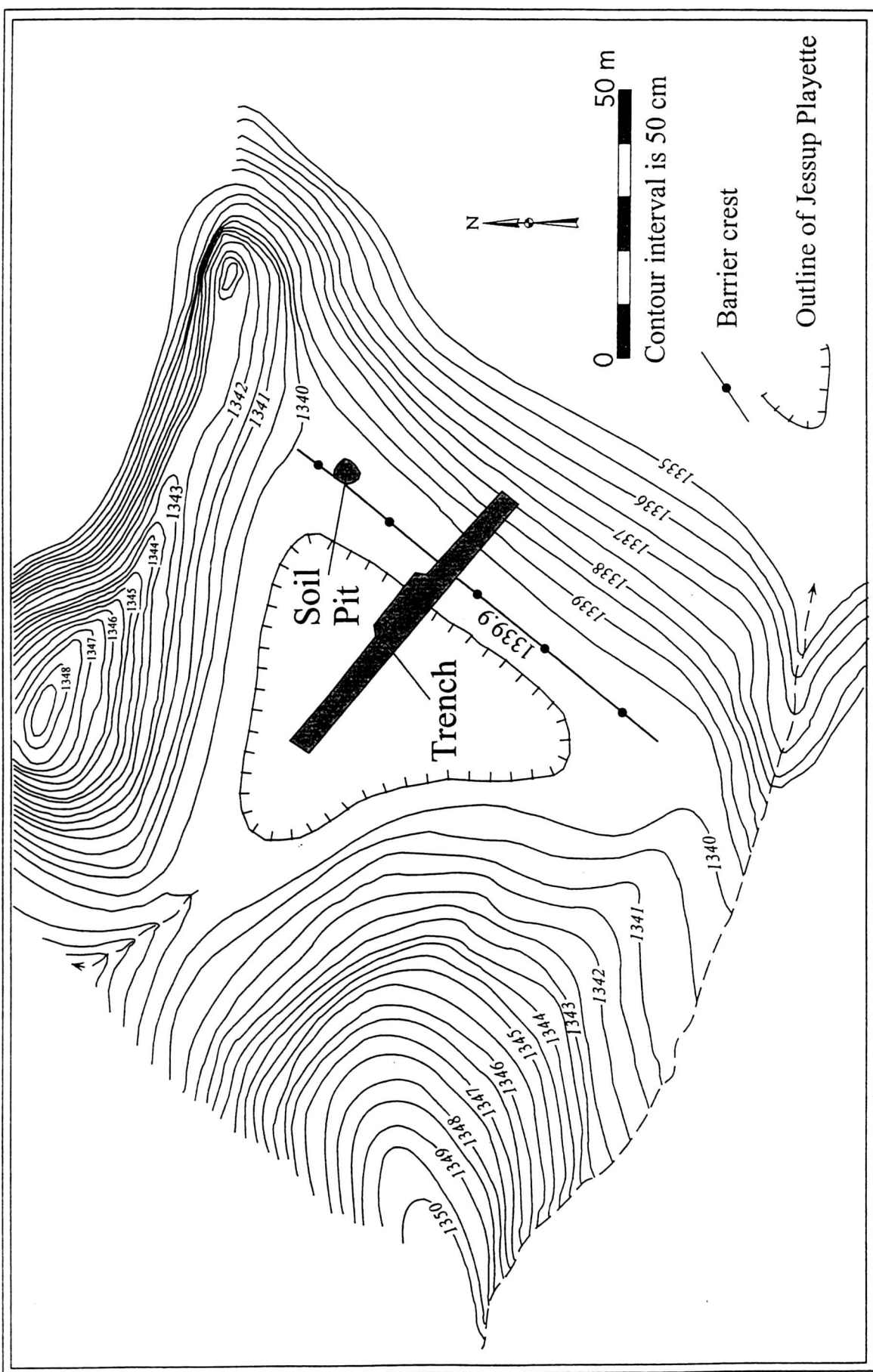


Figure 9. Topographic map of the Jessup Playette trench site showing the location of the trench and soil pit in relation to the highstand barrier and its associated playette. The enclosing barrier marks the highstand of pluvial Lake Lahontan.

Jessup Playette Trench

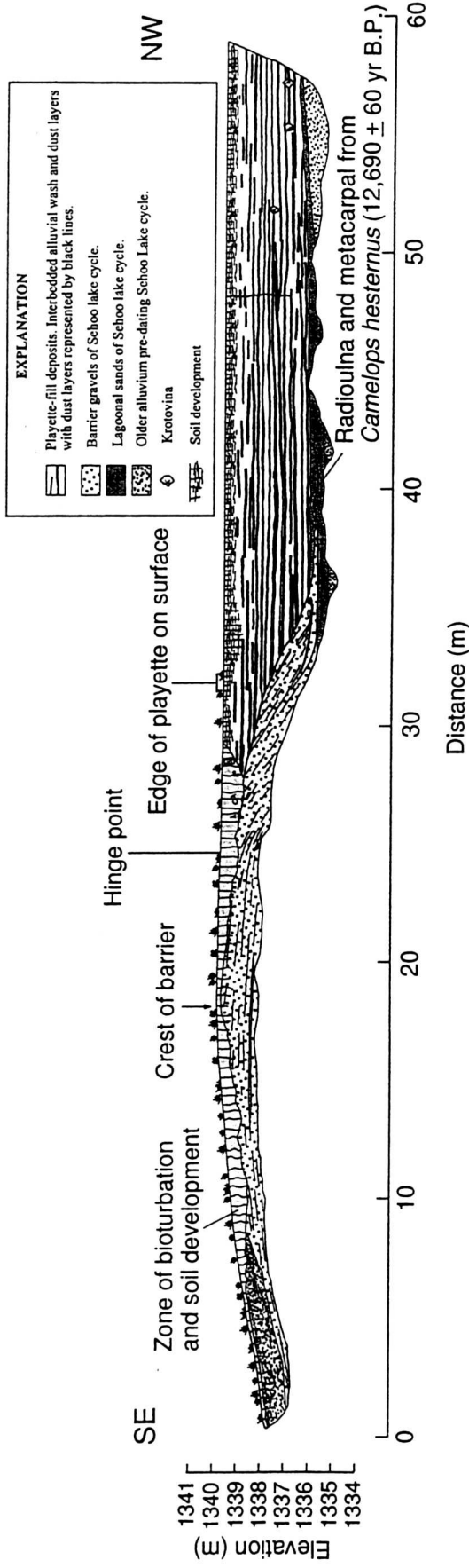


Figure 10. Log of the Jessup Playette trench showing the location of the air-fall tephra and camel bones in relation to the sediments of the barrier, lagoon sands and playette-fill deposits. An AMS 14C date on collagen extracted from the camel radiolna bone yielded an age of 12,690 ± 60 yr BP, thus providing a minimum limiting age on the highstand. The location of the camel bones was projected from the northeast wall onto the southeast wall shown in this log.

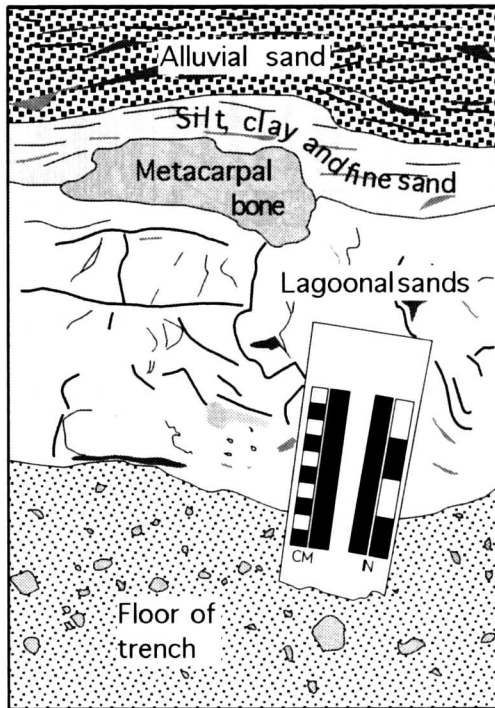


Figure 11. Photograph and drawing of in situ camel bone (12,690 \pm 60 yr BP) in the Jessup playette trench at contact between the lagoonal sands and the playette-fill package.

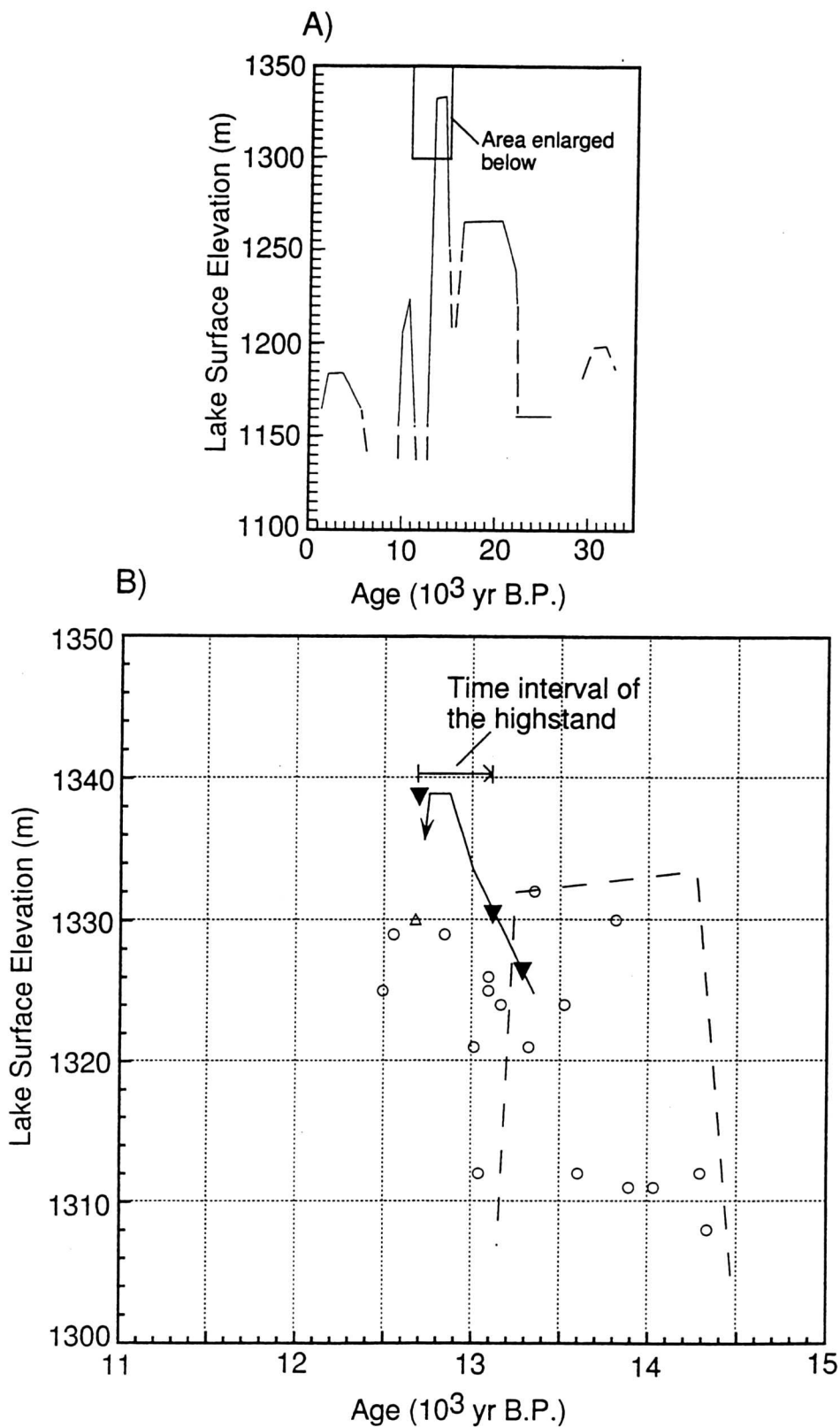


Figure 12. Figure showing Benson et al's (1995) Lahontan lake level curve compared with data and interpretations from this study. A) Simplified lake level curve from Benson et al (1995). B) Lake level curve constructed from the data from this study superimposed on the enlarged lake level curve of Benson et al (1995). Inverted black triangles designate the radiocarbon ages and elevations from this study. Hollow circles represent radiocarbon ages and elevations of tufa samples from Benson et al (1995). Right-side up hollow triangle represents radiocarbon age and elevation of organic material in rock varnish on high terrace from northern Pyramid Lake basin (Dorn et al, 1990).