Genesis and deformation of holocene shoreline terraces, Yellowstone Lake, Wyoming
by Grant Arnold Meyer

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Earth Science
Montana State University
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Abstract:
Geodetic leveling over a 62-year period has revealed uplift averaging up to 15 mm/yr within the 0.6 Ma Yellowstone caldera. This historic uplift is most likely due to magmatic activity at shallow crustal levels, but the time of initiation, history, and significance of uplift is unknown. In this study, postglacial shoreline terraces (dating from ca. 9000 yr B.P. to present) around the north end of Yellowstone Lake are used to measure tilting and estimate net vertical deformation since their formation.

The study area lies within the southeastern caldera rim and contains a sequence of 6 or more discontinuous raised shoreline terraces. 106 profiles were surveyed by accurate leveling techniques across terrace sequences. Shoreline elevations were determined from profiles by extrapolation of the best-preserved wave-cut cliff and platform slopes. Individual terrace segments were correlated across gaps using morphology, vertical spacing, and extrapolation of shoreline tilts. Shoreline, elevation data show that the terraces are significantly and complexly deformed, with local tilt values of up to 6000 microradians (6 m/km). The overall pattern of deformation suggests that inflation somewhat similar to the historic form has occurred throughout the time- of shoreline formation. Uplift rates calculated from radiocarbon-dated shorelines are less than one-half of historic rates, suggesting episodic deformation or reversals in direction which result in lower long-term rates. Though uplift rates may be increasing at present, the terrace deformation does not suggest an overall trend of accelerating uplift.

Many deviations from the historic inflation pattern are observed in the form of strong local deformation, which may be due to local magmatic events such as dike or cupola injection, adjustments to surface distension accompanying inflation, or extension on regional tectonic trends.

Since vertical deformation of the lake outlet area directly affects lake level, it is likely that magma-related deformation has played a major role in controlling lake level fluctuations and terrace formation. Localized tectonic downwarping of the outlet may also have been involved in lake level control. Further work, including extensive absolute dating, is necessary to understand the temporal relationships between individual lake level stands and episodes of deformation.
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YELLOWSTONE LAKE, WYOMING

by
Grant Arnold Meyer

A thesis submitted in partial fulfillment
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June 1986
APPROVAL

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Grant Arnold Meyer

This thesis has been read by each member of the thesis committee and has been found to be satisfactory regarding content, English usage, format, citations, bibliographic style, and consistency, and is ready for submission to the College of Graduate Studies.

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Date  May 5, 1986
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ABSTRACT

Geodetic leveling over a 62-year period has revealed uplift averaging up to 15 mm/yr within the 0.6 Ma Yellowstone caldera. This historic uplift is most likely due to magmatic activity at shallow crustal levels, but the time of initiation, history, and significance of uplift is unknown. In this study, postglacial shoreline terraces (dating from ca. 9000 yr B.P. to present) around the north end of Yellowstone Lake are used to measure tilting and estimate net vertical deformation since their formation.

The study area lies within the southeastern caldera rim and contains a sequence of 6 or more discontinuous raised shoreline terraces. 106 profiles were surveyed by accurate leveling techniques across terrace sequences. Shoreline elevations were determined from profiles by extrapolation of the best-preserved wave-cut cliff and platform slopes. Individual terrace segments were correlated across gaps using morphology, vertical spacing, and extrapolation of shoreline tilts. Shoreline elevation data show that the terraces are significantly and complexly deformed, with local tilt values of up to 6000 microradians (6 m/km). The overall pattern of deformation suggests that inflation somewhat similar to the historic form has occurred throughout the time of shoreline formation. Uplift rates calculated from radiocarbon-dated shorelines are less than one-half of historic rates, suggesting episodic deformation or reversals in direction which result in lower long-term rates. Though uplift rates may be increasing at present, the terrace deformation does not suggest an overall trend of accelerating uplift.

Many deviations from the historic inflation pattern are observed in the form of strong local deformation, which may be due to local magmatic events such as dike or cupola injection, adjustments to surface distension accompanying inflation, or extension on regional tectonic trends.

Since vertical deformation of the lake outlet area directly affects lake level, it is likely that magma-related deformation has played a major role in controlling lake level fluctuations and terrace formation. Localized tectonic downwarping of the outlet may also have been involved in lake level control. Further work, including extensive absolute dating, is necessary to understand the temporal relationships between individual lake level stands and episodes of deformation.
INTRODUCTION

The Nature of the Problem

Surficial geologic mapping by Richmond (1973, 1974, 1977) and Richmond and Pierce (1972) has delineated a series of postglacial shoreline terraces around virtually the entire shore of Yellowstone Lake. The terraces were formed subsequent to the complete disappearance of the Pinedale icecap from the lake basin ca. 12,000 yr B.P. (Richmond, 1976a). They were identified by elevation above present lake level, implying undeformed horizontal shorelines. Radiocarbon dates on four terraces from widely spaced localities over the lake basin are consistent with a nearly constant rate of decline of lake level from about 9000 yr B.P. to present. Lake level decline was ascribed to downcutting of the lake outlet through glacial sediments (Richmond, 1969).

Precise geodetic releveling of benchmarks throughout Yellowstone Park has revealed that uplift comparable in rate to that measured at some active volcanic centers is presently occurring within the 0.6 Ma Yellowstone caldera (Pelton and Smith, 1982) (Fig. 1). First- and second-order level line benchmarks (first established in Yellowstone in 1923) were reobserved to determine relative vertical surface...
Figure 1. Uplift (mm/yr) in the Yellowstone caldera for the period 1923-1975 (modified from Pelton and Smith, 1982). Note the pattern of historic uplift within the study area. The Sour Creek (northeast) and Mallard Lake (southwest) resurgent domes are shaded. Caldera features from Christiansen (1984).
displacements and velocities. These data were developed using a benchmark near the East Entrance of Yellowstone Park as a relative zero point outside of the inferred areas of magmatic and regional tectonic deformation. A map of contoured uplift rates for a 52-year period (Fig. 1) shows a pattern of elongate uplift roughly symmetric about the northeast-trending long axis of the caldera; uplift decreases to zero near the caldera rim. Maximum uplift magnitudes of about 700 mm were reported for benchmarks near the southwest flank of the Sour Creek dome, reflecting an asymmetric rise of the northeastern caldera relative to the southwestern sector. Subsequent releveling (Dzurisin, 1985) and microgravity data (Smith et al., 1984) have confirmed this uplift and measured maximum uplift rates of about 20 mm/yr 5 km north of the outlet of Yellowstone Lake. Relatively high rates of tilting are observed along the northeast shore of the lake. Maintenance of these rates for more than a few hundred years would measurably deform the shoreline terraces. Pelton and Smith (1982) cited a terrace mapped by Richmond (1977) at 18-20 m above the lake at widespread localities as evidence that either (1) uplift began less than 500 years ago, or that (2) despite vertical surface movements in the Holocene, little net deformation has resulted. However, given the discontinuous nature of the terraces, it is possible that unrecognized deformation of shorelines exists.
The primary purpose of this study is to test the hypothesis of horizontality of the Yellowstone Lake terraces and to examine the nature of any deformation which may exist. Shoreline terraces of both marine and lacustrine origin define time-constrained, originally horizontal lines and have been used to determine the magnitude, rate, and style of tectonic deformation, as in studies of marine terraces near the southern California and New Zealand plate margins (e.g., Bradley and Griggs, 1976; Pillans, 1983). They have also been used as records of isostatic uplift, as in studies of raised marine shorelines in the Canadian Arctic (e.g., Andrews, 1970) or the terraces of Pleistocene Lake Bonneville (e.g., Gilbert, 1891; Crittenden, 1963). While Yellowstone Lake is smaller than most lakes where this method has been applied, the essence of the problem lies in the definition of a past water level from terrace morphology and correlation and dating of discontinuous and eroded terrace segments so that a history of quantitative vertical movements can be resolved. The formation of a series of apparently discrete and variably-spaced water level indicators such as these terraces (as opposed to the monotonous succession of beach ridges typical of emerging shorelines in areas of continuous isostatic uplift, e.g., Blake, 1975; Hillaire-Marcel and Fairbridge, 1978) in itself implies quantum water level changes due to episodic tectonic, erosional, depositional, or climatic events;
resolution of these events through study of the shoreline terraces may provide a better picture of the geologic evolution of the Yellowstone Lake area.

Study Area

For reasons of access and relevance to contemporary deformation, the study area was restricted to the northern shore of Yellowstone Lake, from Sand Point on the west to Lake Butte on the east (Fig. 2). All terraces examined lie inside of the topographic rim of the caldera, and with the possible exception of those terraces east of Sedge Creek, within the caldera structural boundary (ring fracture zone) as well (Christiansen, 1984). They are entirely within the area of historic uplift and extend nearly to the zero isoline of uplift as determined by Pelton and Smith (1982). The raised shorelines defined by these terraces lie both nearly parallel to and perpendicular to historic uplift isolines. Terrace development and continuity is reasonably good, and the northwest shore contains a number of paleolagoonal areas important for absolute age-dating of shoreline levels.

Geologic Evolution of the Lake Basin

Yellowstone Lake occupies an approximately 30 km by 25 km, complexly-shaped basin that is the result of multiple processes, including caldera collapse and postcollapse volcanism, ice-cap glaciation, and lacustrine processes in
Figure 2. Map of study area showing localities mentioned in text. Heavy lines = major roads; dashed lines = side roads.
interglacial and postglacial intervals. Three major caldera-forming eruptive cycles have occurred in the Yellowstone region in late Pliocene and Quaternary time, each characterized by long periods of episodic rhyolitic lava extrusion before and after a climactic explosive pyroclastic eruption (Fig. 3) (Christiansen, 1984). Development of the lake basin began during the third cycle with the eruption of the Lava Creek Tuff and the resulting collapse of the Yellowstone caldera at about 0.6 m.y. B.P. Collapse took place on two partly overlapping ring fracture systems, which probably represented two broad culminations in the roof of the large underlying magma chamber. The northeastern ring fracture system transects the present lake area from Sedge Bay to West Thumb Bay and intersects the southwestern ring fracture system in the area north of West Thumb. The dual nature of the caldera is also reflected in the twin resurgent domes, which formed within each ring fracture zone shortly after collapse: the Mallard Lake dome on the southwest, and the Sour Creek dome on the northeast. Each resurgent dome is cut by a complex system of axial graben faults. Although it laps on the edge of the Sour Creek dome, a flow of the postcaldera Plateau Rhyolite that is the same age as the Lava Creek Tuff (within error limits) is not domed, showing that both resurgence and rhyolite lava extrusion began shortly after caldera collapse (Christiansen and Blank, 1972).
Figure 3. Tectonic map of the greater Yellowstone region. Heavy lines mark the rims of the first- (I), second- (II), and third-cycle (III) calderas. Dotted lines mark the outer edges of the resurgent domes within the 0.6 Ma third-cycle Yellowstone caldera. Lighter lines show faults (dominantly normal), with major fault zones (FZ) labeled. F = locality at which Holocene shoreline terraces are offset by normal faulting. Modified from Christiansen (1982).
Paleodrainages corresponding to the Southeast and South Arms of the lake developed as a drainage system into the south side of the caldera (Otis et al., 1977). The Southeast Arm paleocanyon probably extended to at least the present area of Hayden Valley, and was scoured by Pre-Bull Lake glaciation into a deep trough prior to impoundment of the lake. The gross morphology of the remainder of the Yellowstone Lake basin was shaped largely by postcollapse volcanism. Collapse of a subcaldera associated with the eruption of the Shoshone Lake Tuff (later informally renamed tuff of Bluff Point in the West Thumb area; Christiansen, 1974) at roughly 165,000 yr B.P., has been suggested for the formation of the West Thumb basin (Christiansen and Blank, 1972). Eruption of the numerous flows of the Central Plateau Member of the Plateau Rhyolite between 165,000 and 70,000 years ago covered much of the caldera floor north and west of the lake basin and overflowed the western caldera rim. The West Thumb flow (Fig. 4) dammed the Southeast Arm paleocanyon at about 144,000 yr B.P. by abutting the flank of the Sour Creek dome a few kilometers north of Fishing Bridge, resulting in an ancestral Yellowstone Lake (Richmond, 1976a). Continuing eruption of the Plateau Rhyolite lavas and renewed uplift of the Mallard Lake dome formed the broad closure on the north end of the basin which contains the present lake. The Elephant Back flow (K-Ar dated at about 150,000 yr B.P.) overlies the slightly older
Figure 4. Generalized surficial geologic map of the study area. Deposits of shoreline terraces are shown as Pinedale and Holocene lacustrine sand and gravel. Modified from Richmond (1974, 1977).
West Thumb flow in the northwestern study area (Fig. 4). It is cut by a northeast-trending graben fault system that may have formed due to magmatic uplift during later eruptions or intrusions.

The Yellowstone Plateau was the locus for several ice-cap glaciations during the Pleistocene; those occurring before about 150,000 yr B.P. are poorly known (Richmond, 1976b; Pierce, 1979). Advances around 150,000 - 140,000 yr B.P. and 100,000 - 85,000 yr B.P. are generally assigned to the Bull Lake Glaciation, while those dated at >45,000 to 10,000 yr B.P. are assigned to the Pinedale Glaciation, following the traditional nomenclature of Blackwelder (1915) for the Wind River Mountains of Wyoming (Richmond, 1976b; Pierce, 1979; Porter et al., 1983). The Yellowstone Lake basin was a center of ice accumulation during both Bull Lake and Pinedale time (Richmond, 1976a).

Though only locally preserved, significant accumulations of open water lake silts suggest that large bodies of open water existed in the lake basin during interglacial periods of the last 150,000 years. Richmond (1974, 1977) has mapped lake sediments radiocarbon dated at 150,000 and 103,000 yr B.P. and assigned them to pre-Bull Lake interglacial and Bull Lake interstadial periods, respectively. Numerous radiocarbon dates on lake sediments (Richmond, 1976b) show that a Bull Lake/Pinedale interglacial lake existed as late as 45,000 yr B.P.
Although several ice advances and retreats probably occurred during the Pinedale Glaciation (Porter et al., 1983), the next major episode of lake formation commenced with the downwasting of the Pinedale icecap and formation of ice-marginal lakes about 14,500 yr B.P. (Richmond, 1976b). The lake basin was probably fully ice free by about 12,000 years ago.

Previous Work: Shoreline Terraces

Although terraces were described by Richmond (1976a) at up to 470 feet above the present level of Yellowstone Lake, the highest of these features were interpreted to be kame terraces. A widespread terrace at 110 feet (33 m) above the lake shows both ice-contact and wave-cut features as might be formed in association with an ice-marginal lake. Other possible shoreline terraces at 90 and 75-80 feet (27 and 23-24 m) above the lake were recognized to the south of the area of this study but are discontinuous and poorly formed. A large number of terraces were formed around the lake after disappearance of glacial ice from the basin. In the area of this study, Richmond (1974, 1977) mapped postglacial terraces at 60-65, 55, 40-45, 30-35, 25, 15, and 10 feet (18-20, 17, 12-14, 9-10.6, 7.6, 4.5, and 3 m) above the lake.

At the south end of the South Arm (Fig. 1), deltaic sands graded to the level of a 60-65 foot terrace overlie lake silts containing organic matter $^{14}$C-dated at 9060 ± 300
yr B.P. (W-2041) (Richmond, 1974). Richmond (1976a) noted numerous large spits and bars associated with this terrace that are not seen at higher levels and considered it to be the earliest terrace formed after complete deglaciation of the lake basin. However, since outwash of waning glaciers is graded to this level it was assigned to Late Pinedale time. Terraces between 55 and 30 feet (17-9 m) above the lake were suggested to have formed during the Altithermal, and those at 25 feet (7.6 m) and lower during post-Altithermal time (Richmond, 1976a). However, the Canyon Village surficial geologic map (Richmond, 1977) lists terraces from 60 to 45 feet as Late Pinedale and those at 35 feet and below as Holocene.

A minimum age of 5590 ± 250 yr B.P. (W-2286) for a 40 foot (12 m) terrace (Richmond, 1974) was obtained from organic matter in a lagoonal deposit at the south end of the Southeast Arm (Fig. 1) (Baker, 1976). The Squaw Lake hydrothermal explosion (Fig. 4) was believed to have occurred with reduction of confining pressure resulting from lowering of the lake from the 35 foot (11 m) level to the 25 foot (7.6 m) level. Organic-rich clay overlying this deposit was $^{14}$C-dated at 3500 ± 250 yr B.P. (W-2734). Also, the 25 foot terrace is cut on the explosion deposit. On a 10 foot (3 m) terrace on Frank Island (Fig. 1), charcoal from a horizon between beach gravels and overlying colluvium gave a $^{14}$C age of 620 ± 250 yr B.P. (W-1999). The above
dates give both minimum and maximum ages for terrace formation; however, if they are considered accurate then a nearly constant decline of lake level of about 2 mm/yr is indicated. This relationship also depends on the assumption that terraces may be correlated throughout the lake basin by elevation, which is invalid if significant deformation exists.

Richmond (1969) initially suggested that postglacial terrace levels were controlled by erosion of glacial sediments at the lake outlet. In general, later reports (e.g., Richmond, 1976a) did not discuss specific causes for terrace formation. However, it was suggested that the outlet at the time of the 60 foot (18 m) terrace was controlled on indurated silica-cemented interglacial lake sand about 2.4 km north of Fishing Bridge.

**General Geology and Geography**

The present geologic character of the Yellowstone Lake basin is the result of the aforementioned interaction of caldera evolution, ice-cap glaciation, and interglacial lake processes. Bedrock lithologies exposed in the Yellowstone Lake basin are almost entirely Quaternary rhyolitic ash-flow tuffs and lava flows within the caldera, and Eocene intermediate rocks of the Absaroka Volcanic Supergroup and related intrusives outside of the caldera boundary. Minor exposures of Mesozoic sedimentary rocks occur in the southern lake basin. Although rarely exposed there, the
Lava Creek Tuff of the third Yellowstone caldera-forming cycle probably underlies unconsolidated sediments in much of the study area (Christiansen and Blank, 1975; Otis et al., 1977). The postcaldera West Thumb rhyolite flow is exposed west of the Yellowstone River and locally in the wave-cut cliff of the northeastern lakeshore (Fig. 4). The overlying Elephant Back flow is exposed on and to the south of Elephant Back Mountain. Eocene volcanics and intrusives crop out at Lake Butte, which lies on the east rim of the caldera.

Most of the lake basin is covered by surficial deposits of pre-Bull Lake to Holocene age (Richmond and Pierce, 1972; Richmond, 1973, 1974, 1977). Surficial geology of the study area is shown in Figure 4. The upper contact of Holocene and Pinedale lacustrine sand and gravel marks the highest extent of the lake shoreline after retreat of glacial ice from the lake basin proper, as mapped by Richmond (1974, 1977). In general, these deposits thinly mantle Pinedale and older glacial and interglacial sediments on which the postglacial terraces are cut. Although no pre-Pinedale surficial deposits are shown on Figure 4 they commonly underlie terrace deposits and are exposed in the present wave-cut cliff and river-cut scarps. Silt-rich and clay-rich materials of this age are quite compact. In the Mary Bay area, some terraces are cut on hydrothermal explosion deposits of Late Pinedale and Holocene age.
In rare cases, terraces are cut on bedrock. Locally, most notably at Storm Point, Steamboat Point, and Butte Springs (Fig. 2), surficial materials including kame gravels and till have been strongly silica-cemented by hydrothermal activity; in general, terraces are not well developed on these indurated materials.

Soils in the lake basin are generally thin and immature, often consisting of O and A horizons of a few centimeters thickness overlying essentially unaltered parent materials where they are developed on coarser lacustrine sand and gravel. Soils on finer, less well-drained silty and compacted sediments are somewhat thicker; organic-rich horizons reach 0.5 m in paleolagoonal depressions. Little evidence of downward leaching and accumulation of soluble inorganic materials is seen. The effect of soil development on terrace morphology is negligible.

The climate of the Yellowstone Lake basin is a function of the broad positive topographic anomaly of the Yellowstone Plateau and is subalpine in nature, with short, cool summers and long, cold winters; mean annual temperature is 0.4 degrees C (Baker, 1976). The cold water of the lake tends to depress daytime temperatures in the vicinity of the lakeshore in summertime. Average annual precipitation is about 80 cm, much of which falls as snow. Prevailing winds are southwesterly during the ice-free months (June-October; National Park Service, unpublished weather data), thus wave
fetch is greatest on the northeast shore of the lake.

Palynological study of paleolagoons occupying kettles in the Southeast Arm has shown evidence for relatively minor climate changes in the Holocene (Baker, 1976). Conditions became gradually warmer and drier following deglaciation of this locality about 13,000 yr B.P.; this trend culminated in the Altithermal interval (locally about 10,000 to 5,000 yr B.P.), here characterized by a dominantly lodgepole pine forest. Subsequent recurrence of spruce and fir pollen in cores indicates a somewhat cooler and moister climate for the Neoglacial period of the last 5,000 years.

Possible Mechanisms of Deformation

Pelton and Smith (1982) concluded that the most likely cause of historic uplift is magmatic processes at a shallow depth within the caldera. Combined geophysical evidence, including teleseismic P-wave delays, seismicity, gravity, magnetics, and heat flow suggests the presence of a large hot to partially molten silicic body at as little as 5 km depth beneath the caldera (e.g., Eaton et al., 1975; Smith and Braile, 1984). The area of maximum uplift also points toward a shallow zone of low P-wave velocity (Lehman et al., 1982) and a -20 mgal local gravity low that may indicate a 10-50% partial melt under the northeastern caldera rim, centered about 10 km northeast of the Sour Creek dome.

Independent confirmation of contemporary uplift has
been provided by additional leveling, microgravity studies, and analysis of Yellowstone Lake gage data. Continuing detailed geodetic releveling of a NW-SE line across the caldera from Canyon Village to Lake Butte (Dzurisin, 1985) shows a smooth profile of uplift over the caldera axis and indicates that maximum uplift rates have increased from about 14 mm/yr for 1923-1975 (Pelton and Smith, 1982) to over 20 mm/yr for 1976-1984. Microgravity reobservations between 1976 and 1983 show similar results to leveling (Smith et al., 1984). Lake gage records, however, suggest that tilting of the lake basin began in the late 1930s (Hamilton, 1984); if this occurred contemporaneously with uplift of the remainder of the caldera, then uplift has occurred at a nearly constant average rate since initiation. The gage records suggest that uplift may be episodic on a scale of one to several years; this concept is supported by the most recent leveling survey, which showed no significant change in benchmark elevations over a one-year period (D. Dzurisin, written communication, 1985).

A recent summary of historical data has shown that vertical deformation in large, young (<1 Ma) silicic calderas is not unique to Yellowstone (Newhall et al., 1984). The 0.7 Ma Long Valley caldera, California, has experienced similar and higher rates of uplift during periods of inflation in the last several years (Denlinger and Riley, 1984). The Pozzuoli area in the 12 km-wide
Phlegrean Fields caldera of Italy has the longest historic record of deformation within a large caldera. Evidence for submergence and emergence of a Roman marketplace indicates an overall trend of slow deflation in the past 2000 years, which has been interrupted at least three times by rapid inflation events. Uplift during the recent 1982-1984 event averaged 2 mm/day (Berrino et al., 1985). These events have been related to relatively minor and localized eruptive and intrusive activity. This system is much smaller and younger (11 ka) than the Yellowstone caldera and is a trachytic rather than rhyolitic volcanic system.

While the close association of historic uplift with the caldera immediately suggests that magmatic activity is responsible, Pelton and Smith (1982) examined several other possible uplift mechanisms. Of these, tectonic horizontal compressive stress and aseismic creep on a reverse fault were shown to be incompatible with the known stress field in the Yellowstone region. Fault plane solutions and Quaternary normal faults indicate a generally vertical maximum compressive stress. However, coseismic deformation related to this regional extensional regime was accepted as a possible contributor to vertical surface movements. The Yellowstone caldera lies in the apex of an approximately right-angle bend in the Intermountain Seismic Belt, where seismicity along the east-west to northwest trend of the Hebgen Lake fault zone and related faults turns southward to
meet a north-south trending seismic zone (probably related to the Teton fault) that extends into the caldera from the south (Fig. 3). In the southern Yellowstone Lake area, dominantly N-S trending normal faults and focal plane solutions are consistent with east-west extension (Smith and Braile, 1984). A N-S trending fault and associated small graben structures cut Holocene shoreline terraces near Eagle Bay, 10 km south of the study area on the west shore of the lake (Fig. 3) (Richmond, 1974). Though data on present seismicity and focal depth of earthquakes in the Yellowstone region show that brittle failure is reduced in intensity and depth inside the caldera, epicenter locations of 1971-1979 earthquakes (Smith and Braile, 1984) suggest that this fault system and related faults are presently active. Seismicity on this general trend extends northward through the study area. It is therefore necessary to consider extensional tectonic deformation (perhaps modified by the intracaldera rheology) as a possible mechanism of terrace deformation.

Pelton and Smith (1982) viewed glacio-isostatic uplift as a highly unlikely mechanism for historic uplift, in part because any rebound would have commenced with the downwasting of the Pinedale Yellowstone ice cap ca. 20,000 yr B.P., with uplift rates decreasing exponentially from complete deglaciation ca. 12,000 yr B.P. to the present. Rebound may have occurred largely during deglaciation and may have been completed prior to the formation of the first
reasonably continuous postglacial terrace ca. 9000 yr B.P. In addition, the pattern of historic uplift is not consistent with that expected from this mechanism. It is possible, however, that some measurable tilting of older Holocene shoreline terraces may have been produced by differential isostatic uplift. Icecaps greater than 50 km in diameter should induce isostatic depression (Bloom, 1978); the Pinedale Yellowstone icecap was approximately 100 km in diameter, with large outlet glaciers extending to the north in Yellowstone Valley and to the south over the present area of Jackson Lake (Richmond, 1976b; Pierce, 1979). Ice attained a maximum thickness of about 900 m in the lake basin and may have been as much as 600 m thick in the southeastern study area (Richmond, 1976a). By analogy, Pleistocene Lake Bonneville had less than half the maximum depth (330 m), but was approximately 520 km long by 200 km wide. Maximum measured isostatic uplift in response to the removal of this load was about 65 m, with tilting of the Bonneville shoreline of up to 0.6 m/km (600 microradians) (Crittenden, 1963). The water depth of Lake Bonneville was locally quite variable, as was the ice thickness of the Yellowstone icecap; the center of isostatic uplift, however, coincides well with the area of maximum water depth averaged over circles of 40 mile radius (Crittenden, 1963). Boundary effects imposed on isostatic adjustment (i.e., the distribution of the load by an essentially elastic crust
over a viscous subcrust) are likely to be of greater importance for the smaller-area Yellowstone icecap than for Lake Bonneville; however, the high heat flow and resulting shallow depth to ductile conditions in the Yellowstone area may somewhat offset these effects. Given these comparisons, it seems likely that some isostatic adjustment has occurred in response to ice-loading in the Yellowstone Lake basin. Whether or not measurable tilting of shoreline terraces has occurred is dependent on the amount of differential uplift along the relatively short baselines of the lakeshore and the rate of deglaciation of the lake basin.

Thus, any observed tilting of the Yellowstone Lake terraces may be interpreted in terms of magmatic-related deformation, regional extension, and for early Holocene terraces, glacio-isostatic uplift. Separation of the signals of these mechanisms is dependent on their relative magnitudes and clarity of patterns and may be problematic.
METHODS

Shore-Zone Landforms and Raised Shorelines

The use of raised shoreline features in defining past water levels of both marine and lacustrine coasts is well established. The most commonly used features are wave-cut terraces, which are erosional landforms, and beach ridges, which are depositional. An idealized cross-section of an actively eroding shoreline is shown in Figure 5. When such a shoreline is abandoned by a relative lowering of water level, the characteristic tread and riser morphology of a terrace is seen. The wave-cut platform, created mostly by hydraulic action and transport of materials over shore-zone bedrock or non-indurated materials, forms the tread of the terrace. The wave-cut cliff, formed even on coasts of small relief by undercutting and mass wasting, forms the riser.

Most studies of raised terraces have used the shoreline angle (Fig. 5), defined by the intersection of platform and cliff slopes, as a water level indicator. Bradley and Griggs (1976) and Kern (1977) found that the present shoreline angle of California sea coasts is generally within 1 m, and always within 5 m, of mean sea level. Variations are due to wave and tidal irregularities, differential erosion of platform rocks, and other factors. Although
weathering platforms are commonly developed above water level by the solution action of wave spray on high energy marine shorelines, these are not observed at Yellowstone Lake due to limited wave energy and solution potential of shoreline materials.

Yellowstone Lake lacks strong tidal fluctuations and is a much smaller system than marine coasts. However, seasonal water level fluctuation is about 1.5 m, with high water occurring in early July (Denison et al., 1982); also, a seiche of a few centimeters maximum amplitude and period of about one hour occurs (W. L. Hamilton, personal communication, 1984). Profiles of the present shoreline were examined to determine the position and variability of
actively eroding shoreline angles with respect to water level. They were found to lie within about 0.3 m of the mean annual high water level as averaged over 1973 to 1982 from U. S. Geological Survey Water Data Reports (e.g., Denison et al., 1982). Although storm events may cause wave erosion during any of the open water months, effective undercutting of the wave-cut cliff apparently occurs during high water periods. Wave energy at lower levels is mostly dissipated on the ramp of the wave-cut platform before reaching the shoreline, with erosion and sediment transport resulting in regrading of the beach profile (Pethick, 1984).

Profiles of lagoon-forming bars of the present shoreline show that bar crests lie within 1-2 decimeters of the maximum recorded water level, which is about 0.6 m above mean high water (e.g., Denison et al., 1982). Shoreline angles preserved behind these bars are also higher than mean high water, but exhibit more variability. These features are apparently formed during storms accompanied by very high water stages.

**Determination of Shoreline Elevations**

Once a shoreline is abandoned, erosion begins to round the crest of the former wave-cut cliff and colluvium is deposited over the shoreline angle, lessening the slope of the cliff. Other surficial deposits such as eolian sand and small alluvial fans from gully erosion of the riser may also
obscure the platform surface. For this reason most studies of marine terraces have relied on exposures of the shoreline angle in dissected or excavated terraces (Kern, 1977; Pillans, 1983) and location of buried platforms by seismic refraction techniques (Bradley and Griggs, 1976) in order to determine shoreline elevations.

Because actual exposures of the shoreline angle of Yellowstone Lake terraces are rare, and logistical limitations prevented a thorough investigation by geophysical surveys, a method was developed for estimating shoreline angle elevations. 106 topographic profiles were surveyed perpendicular to raised shoreline sequences and plotted in the field (Fig. 6). All profiles were surveyed using a precision automatic level and a 4 m leveling rod. Lake level provided a convenient datum surface for beginning and closing leveling traverses. The rod was held on a stable point on the lake bottom and an average water depth was determined from rod readings of wave minima and maxima, which usually spanned a range of a few centimeters. This figure was then subtracted from the initial backsight reading to give height of instrument above the lake surface. Temporary benchmarks were left at the lakeshore for later releveling if rough wave conditions prevented an accurate water level reading. Since lake level varies by about 1.5 m over the summer field season (e.g., Denison et al., 1982), elevations were later standardized to height above the zero
Figure 6. Terrace profile locations (numbered lines).
mark on the staff gauge at Bridge Bay, using daily visual readings from the gauge (W. L. Hamilton, personal communication, 1984). Elevations reported as "height above datum" in this study refer to this point. Elevations of terraces reported by Richmond (1973, 1974, 1976a, 1977) and Richmond and Pierce (1972) may be converted to this datum by adding an approximate factor of +1.0 m.

Profiles were begun at the water's edge whenever possible. A cloth tape was laid along the profile line and elevations were taken at 1 to 10 m intervals depending on topographic regularity. Because previous work indicated that postglacial shorelines at some localities do not extend above 18-20 m above the lake (Richmond, 1976a), and because surveying is very time-intensive, profiles were not carried higher than terraces with shorelines over 28 m above datum. Bearings of profile lines were taken, and profiles were located by triangulation and topography on 15-minute quadrangles.

All profile lines were closed traverses, either double-run or closed loops. Closures on lake level datum were generally less than 1 cm, indicating that instrumental and surveying error (with the included factor of lake level datum changes over the traverse period due to seiche and lake inflow/outflow variations) is insignificant in measuring shoreline elevations. Minor additional error (+2 cm) is added due to the approximate time calibration of lake
level measurements with Bridge Bay staff gauge readings.

Most terrace treads are broad enough that their lower ends are beyond the main colluvium deposition at the base of the former wave-cut cliff. Also, due to their low slopes (<1 to about 3.5 degrees), little erosional modification of platforms has taken place other than local incision by tributary streams. Therefore, the slope of the lower tread surface as interpreted from the profile and field evidence was projected underneath the colluvium at the cliff-slope base to approximate the original platform surface (Fig. 7, Fig. 8). The former cliff slope was approximated by projecting the steepest remaining slope segment of the terrace riser downwards to intersect the tread projection. This intersection gives a reasonable reconstruction of the original shoreline angle and its present elevation for the relatively young, small terraces of Yellowstone Lake. Although the steepest slope of the terrace riser only roughly approximates the original cliff-slope angle, a relatively small amount of error is introduced by this method due to the low slope angle of the platform surface. The location of the platform surface is more critical to determination of the shoreline elevation. Although beach deposits cover the actual eroded platform surface, they are graded to a minor thickness at the shoreline (generally, the underlying material is exposed in a patchy manner) and thus their surface can be used as a substitute. With the
Figure 7. Interpretation of shoreline angle elevations from a surveyed profile of terraces with well-developed morphology. Short dashed lines show wave-cut cliff and platform extrapolations. Solid circles mark interpreted shoreline angle elevations, with range of uncertainty shown by error bars.

Figure 8. Interpretation of shoreline angle elevations from poorly developed terrace morphology. Conventions same as in Figure 6. Interpretations here were based partly on lateral continuity with well-developed terraces. Open circles indicate that wave-cut origin of the slope break is uncertain.
exception of the cliff-derived colluvial wedge and local accumulations of eolian sand, deposition over the beach deposits has been insignificant, generally less than a few centimeters of humus and organic soil. Possible error introduced by approximate reconstruction of platform and cliff slopes is generally less than ±0.4 m. Larger errors are possible where terrace morphology is poorly expressed.

Raised lagoonal shorelines present a somewhat different morphology from which to interpret shoreline elevation (Fig. 9). In most cases, some degree of wave-cut shoreline morphology was formed prior to closure of the lagoon by the barrier bar. Also, the top of the bar represents a beach ridge crest formed by a storm-induced maximum water level. Although tread and riser morphology behind the bar crest is often subtly expressed, shoreline elevations determined from these features are generally within a 1-2 decimeters of the elevation of the bar crest. Due to the storm-related nature of these features, some systematic error toward higher elevations may be incurred by use of lagoonal shorelines. However, despite the fact that larger error brackets are assigned to data points from lagoonal shorelines, large deviations are not apparent in the elevation trends of the shoreline segments in which they are included.

The slope angle of platform surfaces (as approximated by terrace tread surfaces) appears to depend mostly on the initial topographic slope of the coast and the degree of
Figure 9. Interpretation of shoreline elevation from a raised lagoonal profile. Conventions same as in Figure 7. Large range of uncertainty is due to poorly defined cliff and platform slopes. Note the relationship of the interpreted shoreline elevation to the eroded bar crest.
development of the terrace. Broad platforms eroded on gentle topography have slopes of 1 degree or less. Steep coastal segments display relatively short, steep platform surfaces (2.0-3.5 degrees). This relationship is expected given the more rapid frictional energy loss and resulting rapid shallowing of wave base on a steep shoreline. The resistance of coastal materials to wave erosion is a significant factor in determining terrace development and platform width. In general, terraces are poorly expressed on indurated materials such as unaltered bedrock and hydrothermally-cemented surficial materials. They are best expressed on cohesive unconsolidated materials such as compact lacustrine silts.

Significant accumulations of eolian sand are found on terrace platforms in the Fishing Bridge Campground, Pelican Creek, Squaw Lake, and Stevenson Island areas (Fig. 2, Fig. 4). These deposits were recognized but generally not mapped by Richmond (1976a, 1977). In a few locations, dunes obscure the shoreline area of a terrace, but with the exception of Stevenson Island dune fields are not extensive enough to seriously affect data collection.

**Correlation of Shorelines**

Terraces around the north end of Yellowstone Lake exist as discontinuous segments. Major headlands including Lake Butte, Steamboat Point, and Storm Point (Fig. 2) lack well-
developed terraces due to resistant rock. Terraces are poorly developed on coastlines of very low relief (as in parts of the Pelican Creek area) and in areas sheltered from storm waves by headlands (e.g., lower terraces in the interior of Bridge Bay). Many small tributary streams have incised terraces, and the lateral migration and downcutting of Pelican Creek has removed all raised shoreline features from a 0.8 km wide swath. Perhaps most significantly, erosion of younger wave-cut cliffs has totally removed higher terraces in many places. An actively eroding wave-cut cliff exists intermittently over more than half of the study area lakeshore; much of this cliff is over 8 m in height, and lower terraces are poorly represented as a result.

The continuity of individual terraces was initially established by walking out shorelines, i.e., the break in slope of the terrace at the buried shoreline angle. A surveying altimeter was used to ensure that a constant elevation was maintained while crossing stream gullies and other narrow terrace breaks. Shorelines were mapped on U. S. Geological Survey 1:62,500 scale 15-minute quadrangles; although this scale map was too small for accurate location, mapping served to record the continuity of segments.

In some cases, unambiguous correlation was possible by extrapolation of tilt or shoreline slope over short distances (generally <0.5 km). This could be done with
confidence only where the sequence and spacing of terraces on either side of the gap resulted in unambiguous matching of shorelines. The same vertical sequence of shorelines cannot be assumed to occur in all areas; a terrace may be absent due to a local break in original development or later removal by erosion. In addition, spacing between any two terraces may vary from place to place due to differences in cumulative deformation.

Radiocarbon Dating

A major concern of this study is absolute dating of shorelines. The search for datable organic materials and volcanic ash was focused on paleolagoonal depressions. Both past and present lagoons are found throughout the lake basin, but are much more abundant on western (generally leeward) shores of the lake. Most were formed by the closure of a small bay or coastal reentrant by a longshore-drift-deposited barrier bar. Eight soil pits in raised lagoons showed that subsequent to closure, about 0.5 m of silty, generally fining-upward lagoon-fill sediments were deposited over the older wave-cut surface which forms the bottom of the lagoon. Little aquatic vegetation can be seen in present lagoons, and very little organic material is present in the silts. The lagoon-fill sediments are sharply overlain by an organic-rich soil which is built up by growth and decay of grass and sedge after abandonment of the
shoreline; this soil horizon is dark brown-black at its base and becomes progressively lighter and less compact upwards to the present humus layer at the surface. The base of this horizon represents the time of abandonment of the shoreline. No volcanic ash-bearing horizons were identified in soil pits.

Ideally, samples for radiocarbon dating should lie near the base of the lagoon-fill sediments and consist of wood, charcoal, or other in situ organic material not transported to this level. However, no material of this sort was found, either in soil pits or in numerous borings of lagoonal depressions with a bucket auger. Since no suitable material was available from lower levels, samples for $^{14}C$ dates were obtained from the base of the organic horizon described above. This horizon lies at shallow depth and is not isolated from later rootlet intrusion and decay, thus is probably contaminated by younger carbon. Also, a significant amount of time, perhaps a few hundred years, is required for the original buildup of decayed material in the abandoned lagoon. Therefore, dates obtained from this basal material must be considered minimum ages for shoreline abandonment.
RESULTS

Projection lines for shoreline elevation data (Fig. 10) were drawn to approximate the trend of the modern shoreline of the lake. The data are shown in Figures 11 and 12 as projected perpendicular to these lines. Shoreline elevations were plotted to the nearest decimeter, which is the limit of resolution for interpretation of profiles. Solid circles indicate good raised shoreline morphology; open circles indicate poorly developed terrace-like features whose wave-cut origin is uncertain. Error bars on shoreline elevations are the sum of the ±0.3 m shoreline angle/water level uncertainty and the variable uncertainty in interpretation of shoreline angle position from the surveyed profile. Data points are sometimes not centered within the range of error because a most likely interpretation falls at the indicated point. The error bars are thus an approximation of interpretive confidence rather than absolute measurement error.

Solid lines between data points indicate correlations established by walking out shorelines. Dashed lines indicate probable correlations established by short distance extrapolation of tilt. The raised shorelines are mostly parallel to subparallel to the projection line segments, thus a close approximation of the true slope or tilt of each
Figure 10. Lines to which shoreline elevation data are projected in Figures 11 through 14 (A-F, I-V). Dashed lines are approximately located contours on historic uplift rates in mm/yr (Pelton and Smith, 1982). Abbreviations used: SA = Sand Point; GP = Gull Point; LH = Lake Hotel; FB = Fishing Bridge; SP = Storm Point; ST = Steamboat Point; LB = Lake Butte (on caldera rim).
Figure 11. Shoreline elevation data for straight-line segments of the northwestern lakeshore (lines I-V on Fig. 10). Trends of each segment are shown. See text for explanation of data points and correlation lines. Shaded regions represent areas of erosion by the present wave-cut cliff and tributary streams. Sample locations for radiocarbon dates are shown.
Figure 12. Shoreline elevation data for the northeastern lakeshore (lines A-F on Figure 10). Conventions and abbreviations same as in Figure 11.
shoreline is shown. The shoreline tilts represent the slope component of the actual deformed surface along the trend of the projection segment. The shoreline elevation diagrams represent three-dimensional data along a single plane, therefore apparent changes in tilt directions between segments may be dominantly the result of different trends of projections of the same surface. For example, the apparent reversal in tilt direction of the 12 m shoreline at Gull Point (between segments III and IV) is most likely due to the 70 degree change in projection. The data for that shoreline fit an essentially planar deformed surface that slopes down to the northeast, though nonplanar solutions are also possible.

Also shown on these diagrams are radiocarbon dates of this study, with arrows indicating sample locations as projected in the same manner as elevations. Samples from raised lagoons at 17.6 m and 9.1 m in the area north of Lake Lodge gave $^{14}C$ ages of 2495 ±135 yr B.P. (GX-10751) and 1410 ±160 yr B.P. (GX-10750) respectively. As discussed previously, these dates give minimum ages for abandonment of each shoreline.

One $^{14}C$ date (GX-10935) is not shown because of obvious discordance. Although lying at a similar stratigraphic level as the above samples, charcoal from the basal organic horizon of a lagoon at 6.6 m above datum south of Gull Point gave an age of recent (-15 ±155 yr B.P.). This material,
which lay at only 20 cm depth, may have been a tree root that was combusted in situ during an intense forest fire.

Shaded regions represent areas of presently active wave-cut cliff and tributary stream erosion. No terraces are preserved within these areas, and correlation of terrace segments across such laterally extensive breaks is difficult.

The data appear to represent at least seven individual terrace levels, if the vertical sequence in segment I near the lake outlet may be taken as representative. It cannot be assumed, however, that the same vertical sequence of shorelines will occur in all areas, as a terrace may be locally absent due either to lack of original development or later removal by erosion. Also, if vertical surface deformation has occurred, the vertical spacing between terraces may vary from place to place due to differences in net deformation over the time of terrace formation.

It is readily apparent from Figures 11 and 12 that the terraces are in many areas tilted well beyond the limits of the error bars for horizontal, undeformed shorelines. The northeast shore in particular shows significant tilting of most well-correlated terrace segments. Averaged local tilting of up to 6000 microradians (6 m/km) is seen in the Fishing Bridge area (segment A). Other terrace segments showing obvious deformation are seen in shoreline segments C (Mary Bay) and F (Sedge Bay), with local tilts to 3000
microradians, and in segment III (Gull Point - Bridge Bay) with tilts of about 1200 microradians.

Despite evidence that shoreline angles of modern lagoons lie well above (about 0.6 m) those of open coasts, no systematic variation is seen between elevations obtained from raised lagoon profiles and those of contiguous, correlative open coastal terraces. This may be because shoreline angles in modern lagoons which represent storm levels are weakly developed but recognizable, while those of raised lagoons are rounded by slope evolution processes. This results in profiles which reflect the larger pre-lagoonal platform and cliff and give elevations similar to open coastal terraces. Also, the bar crests of raised lagoons do not lie significantly higher than correlative shoreline angles. This may be expected, since the cohesionless, exposed bar gravels would be quickly lowered in profile when the bar became inactive.

The shoreline elevation data show apparent localized warping and folding that is significant within error limits and deviates from larger tilting trends, particularly in the Mary Bay and Sedge Bay areas (segments C and F). It is possible that some of this variation is due to interpretation problems or other data noise of indeterminate origin. However, in segment C, a downfolding pattern with a wavelength of roughly 1.5 km is consistent through at least four shorelines, lending support to a deformation-related
No faults have been mapped which cut Pinedale or younger deposits in the study area (Fig. 4), and no geologic or geomorphic evidence for fault offset of terraces was noted in this study. However, the possibility of Holocene-age faulting still exists; where shoreline elevation data most strongly suggests faulting, the data set is incomplete and inconclusive. For example, the folded set of terraces in segment C may be offset across a gully which lies at the intersection of segments C and D, but terraces are so poorly preserved within D as to make this hypothesis untestable. Similarly, evidence of faulting associated with the strongly tilted terraces in the Fishing Bridge-lake outlet area may have been removed through erosion by the Yellowstone River.

Although crosscutting of older terraces by younger ones was not directly observed, it may occur in two areas. On either side of the lake outlet at Fishing Bridge, higher shorelines (between 20 and 6 m) are strongly downtilted toward the Yellowstone River so that their trends converge with the younger, nearly horizontal shoreline(s) at 5.5 to 6.0 meters. Unfortunately, the gap eroded by the river precludes clear understanding of correlations and relationships. At Sedge Bay, between Steamboat Point and the caldera rim at Lake Butte, a terrace falls from 5.5 to 3.1 m above datum and appears to be crosscut by a nearly horizontal terrace segment at 2.3 m. It is possible,
however, that the upper segment may correlate southeast across the mouth of Sedge Creek to the nearly horizontal segment at about 4.5 m.
As a first approximation, correlations were drawn between terrace segments using terrace morphology, extrapolation of shoreline tilts, and vertical spacing (Figs. 13 and 14). Separately these criteria are unreliable for correlation. Terrace morphology or the degree of development of a shoreline is strongly dependent on local factors such as the material being eroded and the orientation of the shoreline with respect to wave radiation energy. For example, while the terrace of shoreline S5 in segment IV is strongly developed (i.e., it has a broad platform and considerable relief on the riser) and continuous, it completely loses expression in segment III as the sheltered area west of Gull Point is entered. On the north and west shores of Bridge Bay, where terrace development is somewhat stronger, this terrace has probably been eroded through during formation of lower terraces.

Correlation by extrapolation of tilt becomes unreliable over distances longer than about 0.5 km due to the close vertical spacing and small-scale warping of terraces in many areas. Vertical spacing of shorelines is often ambiguous in establishing correlation, since at any given location some
Figure 13. Tentative correlations (dotted lines) for raised shorelines of the northwestern lakeshore (Fig. 10). Conventions as in Figure 11. Shorelines are numbered arbitrarily from lowest to highest (S1-S8). Locations of radiocarbon minimum ages for S5 and S6 are shown.
Figure 14. Tentative correlations (dotted lines) for raised shorelines of the northeastern lakeshore (Fig. 10). Conventions as in Figure 12 and 13.
terraces of the sequence may be missing due to either lack of original formation or later erosion. The absolute spacing of terraces over the study area will vary if any nonuniform vertical surface movements have occurred over the time of formation, and relative spacing will vary if the deformation pattern has changed over time.

However, in combination these criteria (along with a general sense of reasonable deformation patterns) do allow some support of correlations. As interpreted, the data represent 8 distinct shorelines, with one higher shoreline measured in segment F. These are numbered from lowest (S1) to highest (S8). No age connotation is implied by numbering, as some evidence (discussed later) suggests that reversals in overall lake level decline may have occurred. No terraces higher than 28 m above datum were surveyed, although postglacial terraces may extend above that elevation in the Pelican Creek area and other locations. However, an unsurveyed terrace that lies several meters above 28 m in the Sand Point area contains kettles and is probably of late Pinedale (ca. 12,000 yr B.P.) age. It must be remembered that the correlations shown in Figures 13 and 14 comprise only one of several reasonable interpretations which should preferably be tested by independent criteria.

**Independent Support of Correlations**

Since accuracy in determination of magnitudes and rates
of vertical deformation is dependent on reliable correlations between terrace segments, various age-dependent characteristics of the terraces were investigated in an attempt to provide more confidence in correlations. Correlation was attempted by radiocarbon dating of lagoonal deposits and by age-dependent aspects of terrace riser slopes. Soil development was initially considered as a correlation tool, but the slow rates of soil development on terrace surfaces discouraged further investigation of the method.

Although radiocarbon dating is potentially the most positive method of establishing correlations between discontinuous terrace segments, two factors prevented effective application in this study. The large number of individual terrace segments requires that an equally large number of samples be dated, at a cost well beyond the limits of available funding. Also, the availability of datable organic material in an appropriate, isolated stratigraphic context was found to be poor. Ideally, samples for radiocarbon dating from lagoons should lie near the base of lagoon-fill sediments and consist of wood, charcoal, or other in situ material which resists contamination. However, no datable material was found at this stratigraphic level. The dated basal organic material was taken from a level corresponding to the abandonment of the shoreline and probably took a significant amount of time to accumulate. In
addition, the dated horizon was not isolated from later rootlet intrusion and decay, thus these dates give only minimum ages for shoreline abandonment. Correlation using such poorly constrained dates is not possible.

Analysis of terrace morphology and the state of erosion of the original shoreline features has potential as a means of correlation. The maximum angle of the riser slope of a shoreline terrace (sometimes referred to as a shoreline scarp; e.g. Bucknam and Anderson, 1979) in a state of erosional degradation is mainly a function of initial slope angle, height of slope, and time since formation (specifically since shoreline abandonment and the end of active undercutting). The initial slope angle may be assumed to be the angle of repose for the slope-forming material and thus essentially the same for similar materials (Nash, 1984). It must be cautioned, however, that large differences in erosional processes and rates can be expected between cohesionless sands and gravels (which are the most suitable for slope degradation analysis; Nash, 1984) and cohesive materials with large proportions of silt and clay.

Bucknam and Anderson (1979) plotted slope angle vs. slope height for Lake Bonneville shoreline scarps and morphologically similar, but younger, fault scarps. All of these scarps were cut on unconsolidated sand and gravel. From these data they inferred a logarithmic relationship between these variables; data from time-correlative scarps
plot as a curve, where the slope of the curve decreases with the age of the scarp. The standard form for this analysis is now semilogarithmic plots of maximum scarp slope angle ($\theta_s$) and scarp height ($H$), where data from correlative shoreline or fault scarps regress to straight lines according to the relation

$$\theta_s = C_1 + C_2 \log(H)$$

where $C_1$ and $C_2$ are constants (Hanks et al., 1984).

Some aspects of the Yellowstone Lake terraces are favorable for this type of correlation, including the wide range of slope heights of terrace risers and the relative youth of the terraces. Also, a large body of morphologic data was available from terrace profiling.

Since terraces are gently tilted and subhorizontal over much of the northwestern lakeshore, the tentative correlations shown in Figure 13 are more reliable than those on the northeastern lakeshore (Fig. 14). Also, variations in slope-forming materials and overall coastal morphology are somewhat less than on the northwestern shore. Therefore, morphologic data from shorelines S2 through S6 on the northwestern shore were examined on a semilogarithmic riser slope vs. riser height plot (Fig. 15). Unfortunately, no diagnostic trends were revealed which are useful in correlation. Confidence intervals for regressions on S2 through S6 (not shown) largely overlap. The slopes of lines for S2, S5, and S6 are essentially similar, and the expected
Figure 15. Semilogarithmic plot of maximum riser slope angle ($\theta_s$) vs. riser slope height ($H$) for shorelines S2 through S6 of Figure 13, with regression lines for each shoreline data set. Line slopes are similar, and confidence intervals for regression lines (not shown) largely overlap, so that shorelines may not be correlated by this method. Coefficients of determination ($r^2$) are given at upper left.
relationship of higher slopes for regressions on lower, presumably younger shorelines is not seen. This may be in part because lower terraces are commonly cut on cohesionless beach sand and gravel of higher lake levels, while higher terraces are more often cut on compact silt-rich till and other fine-grained glacial and lacustrine sediments. As a result, lower terraces may have lower initial slope values and degrade more rapidly. The trend of lower coefficients of determination ($r^2$) for higher terraces may occur partly because slopes underlain by cohesive materials are not expected to behave as predicted by this method.

Other difficulties prevent practical application of this method. Most segments of positively correlated shoreline are of limited length, thus have limited numbers of data points and often little variation in slope heights. As a result data sets are composed of points of insufficient number and spacing to confidently define lines, unless assumptions of correlations (as in Figures 13 and 14) are first made. Also, the age difference between adjacent terraces is probably small relative to their absolute ages, so that even with uniform materials poor separation may exist between lines.

A possible method of correlation and dating not explored in this study is obsidian hydration dating. The thickness of the hydration rind on an obsidian surface (which may be the wall of a microfracture) is a function of
time and temperature since exposure of the surface, and composition of the glass (Friedman and Long, 1976). Critical factors in applying this method to dating of beach gravels associated with terraces are: 1) identification of cracks or surfaces which were "fresh" or unhydrated at the time of deposition; 2) location of sample points at sufficient depth (1.5 to 3 m minimum; Pierce et al., 1976) and away from anomalous heat flow such that a reliable and consistent integrated thermal history can be assumed for each sample; and 3) collection and dating of enough samples per terrace segment to allow statistically valid correlation of segments. Observation of modern beach gravels should indicate if hydration rinds are removed in the beach environment.

Significance and Correlation of Radiocarbon Dates

If Holocene vertical deformation in the Yellowstone Lake basin is assumed to be restricted to within the caldera, then dated terraces outside of the caldera may be used to develop a lake level history relative to a stable datum. For discussion purposes, all dates are here considered as accurate ages of terraces, although none can actually be interpreted to precisely date the time of a lake level stand. No historic geodetic measurements have been made along the lakeshore outside of the caldera, and active faults in the southern lake basin (Fig. 2) may have
significantly deformed shorelines. Thus terraces measured and dated by Richmond (1976a) at sites outside of or just within the caldera rim are here used to construct only a tentative lake level history. These dates are plotted against their shoreline elevations in Figure 16. They suggest a linear decline of lake level of about 2 mm/yr since ca. 9000 yr B.P.

Dates correlated with shorelines S5 and S6 are also plotted on Figure 16. These dates were obtained from adjacent sites north of Lake Lodge, where historic uplift has been approximately 10 mm/yr. They plot well above the "stable" lake level line and suggest emergence of shorelines at this locality at a rate of 6 mm/yr. About 4 mm/yr of actual uplift is thus indicated. However, much of this apparent uplift may be the result of minimum ages for S5 and S6 which reflect younger carbon contamination and the time required for organic accumulation after shoreline abandonment. Other evidence suggests that the linear rates of uplift and lake level decline implied by Figure 16 are greatly oversimplified average rates; this evidence is discussed later in this chapter, under "Controls on Lake Level and Terrace Formation".

**Relationship of Terrace Deformation to Historic Uplift**

The presently active inflation of the caldera floor suggests a similar mechanism for Holocene terrace
Figure 16. Height of shoreline vs. $^{14}$C age of terraces. Solid line approximates decline of lake level relative to a stable datum using dates reported by Richmond (1976a) for terraces outside of the caldera (W-2041, W-2286) or just within the caldera rim (W-1999). Minimum ages for terraces obtained in this study (GX-10750, GX-10751; dashed line) suggest a more rapid rate of emergence of shorelines north of Lake Lodge due to uplift within the caldera and/or radiocarbon ages that are too young.
deformation. Therefore, the terrace deformation was examined for similarities in pattern to historic uplift. In a very general sense, the terrace deformation (Figs. 13 and 14) is consistent with uplift of the caldera axis (Fig. 1, Fig. 10): excluding the downwarping at Fishing Bridge, shorelines above S2 on the northeastern lakeshore show a net rise to the northwest, toward the center of the caldera. With minor exceptions, higher shorelines show greater amounts of tilting, suggesting cumulative uplift over the entire time of terrace formation. Shorelines on the northwest shore, whose trends are nearly parallel to the uplift contours of Pelton and Smith (1982) (Fig. 10), show little net change in elevation. A uniform pattern of deformation requires that a relatively small amount of differential uplift should have occurred between segment V and segment I. However, for most of these shorelines any significant elevation change is down to the north, even excluding the outlet area downwarping. It should be noted that the density of data on historic uplift along the northwest shore is relatively low (Pelton and Smith, 1982), such that the local pattern of uplift is poorly resolved.

The lowest shoreline (S1), although discontinuous, appears to be horizontal, lying everywhere at 2.3 ±0.2 m above datum. This is consistent with the interpretation of this level (the 5 foot terrace of Richmond, 1974) as a "storm beach" which is preserved above normal wave erosion.
This shoreline was measured at protected sites, predominantly behind lagoon-forming bars, where its mean elevation is within 0.1 m of the maximum recorded water level of the lake (Denison et al., 1982). It is apparently young enough to be undeformed within the limits of the shoreline elevation method.

Shoreline S2 shows a gentle rise in elevation toward the lake outlet along both shores, and like S1, shows no downwarping at the lake outlet. It rises more rapidly along the northeast shore than the northwest and thus is the most similar shoreline in deformation pattern to historic uplift. The magnitude of differential uplift of S2 is consistent with a few hundred years of uplift at historic rates.

Shorelines S3 through S8 all show net differential uplift toward the caldera axis, but strong local downwarping is also seen. Each higher shoreline shows somewhat greater uplift than those below it, supporting the concept of increasing age of shorelines with elevation. Uplift magnitudes suggest duration of uplift at historic rates which range from roughly 300 years for S3 to 700 years for S6. Shorelines S7 and S8 are poorly represented by these data, but tilting directions of their included shoreline segments are mostly congruent with lower shorelines. The above magnitudes are based on correlations which result in conservative estimates of uplift. It should be noted that other reasonable sets of correlations result in uplift
magnitudes which may be as much as 100% greater for some shorelines. It is unlikely that net uplift of shorelines is much less than estimated here.

Because shoreline S6 contains the longest segment of well-correlated terraces on the northeastern lakeshore and is dated (albeit by correlation to a terrace on the northwestern shore), it was used to calculate a net tilting rate. In segment C, which is roughly perpendicular to historic uplift isolines (Fig. 10), averaged tilting of S6 of about 1750 microradians over a minimum of 2500 years since terrace formation gives a maximum tilting rate of about 0.7 microradians/yr. This is about 2/3 of the historic (1923-1975) rate along this shoreline. Tilting rates are thus lower over longer time periods (8 yr rate > 52 yr rate > 2500 yr rate). This trend probably reflects episodicity or reversals in direction of deformation and is commonly noted in measurements of tectonic deformation as well (Schumm, 1963).

If the caldera rim (at the east end of segment F, Figure 14) is taken to be a stable base at which no deformation has occurred, then the absolute magnitude of uplift of S6 through segment C approaches 10 m, using the tentative correlations. This estimate gives an maximum uplift rate of 4 mm/yr, as compared to the historic rate of 7.5 mm/yr along this part of the lakeshore. Net uplift of this shoreline over the entire northeast shore is unknown,
as no elevation data were collected as it approaches the lake outlet. The elevation of the correlated shoreline (S6) on the west side of the outlet, however, suggests that net uplift may not exceed 10 m. This would mean that the maximum uplift rate over the northeast shore is less than 1/2 of the historic rate (10 mm/yr from Lake Butte to Fishing Bridge; Pelton and Smith, 1982).

Causes of Terrace Deformation

No firm conclusions may be drawn on the origin of terrace deformation based solely on tentative correlations, but most reasonable sets of correlations support uplift of the caldera interior somewhat similar in form to historic deformation. As discussed previously, several lines of geologic and geophysical evidence suggest that the historic uplift is due to magmatic activity at shallow crustal levels within the Yellowstone caldera. In the absence of conflicting evidence, it may be assumed that the terrace deformation as well results mostly from magma-induced surface movements.

Evidence from other young ash flow calderas indicates that ongoing magma-related ground deformation is common. Of the few well-studied analogs, the 0.7 Ma Long Valley caldera, California, is probably closest to the Yellowstone system in terms of present state of activity. It has undergone similar and higher rates of inflation accompanied
by intense seismicity during the period 1979-1985, with maximum cumulative uplift of 0.5 m (Hill et al., 1985). Maximum cumulative uplift at Yellowstone is 0.9 m between 1923 and 1985 (Dzurisin, 1985). The form of uplift at Long Valley is relatively well constrained with both vertical and horizontal data; in general it is radially asymmetric and centered on the resurgent dome, with diffuse right lateral strike-slip motion in the south moat of the caldera. The most recent source models for this deformation incorporate seismic shear wave studies to interpret expansion of a shallow protrusion or cupola above the main residual magma chamber, with possible sill-like expansion of the top of the magma chamber (Denlinger et al., 1985). Some models suggest that unusual seismicity in the south moat reflects the intrusions of a series of dikes (Hill et al., 1985).

Where the roots of older large calderas have been exposed by erosion, evidence of extensive intrusive activity during the first several million years following collapse is noted (Lipman, 1984). Commonly seen above a large subcaldera granitic pluton are dikes and small resurgent plutons or cupolas intruding the shallow crust. Dikes are especially common along ring fractures, with some occurring along adjoining zones of extension outside of the caldera. A few dikes were feeders for eruptions. While historic uplift at Yellowstone shows a relatively uniform pattern of inflation, it is possible that some shoreline
deformation results from more localized magmatic activity as in the examples above.

The role of glacio-isostatic deformation in shoreline deformation remains unknown. Those shorelines with sufficient length of correlation to resolve tilting suggestive of isostatic rebound are almost certainly too young to have been significantly tilted, and the overprint of presumed magma-related deformation almost certainly masks the signal of rebound. Older Holocene shorelines outside of the caldera present the best opportunity for resolving this question.

While no faulting of terraces was directly observed, some features of terrace deformation suggest tectonic regional extension as a possible cause. Since evaluation of this mechanism requires a more detailed discussion of deformation patterns it is discussed below.

Complications of the Deformation Pattern

The deformation pattern of the Yellowstone Lake shorelines is significantly more complex than the relatively smooth profiles of historic uplift. As suggested by older caldera systems, it may reflect numerous episodes of magma-related deformation of varying form. It appears to indicate long-term cumulative inflation of the caldera, but uncertainty in correlations prevents strong support of this conclusion. It must also be remembered that while a
significant portion of the caldera floor is represented by
the study area shorelines, they still provide only a partial
picture of the deformational history of the caldera and are
subject to dominant influence by large, local deformation
events.

With all sets of correlations, the largest deviation
from the historic uplift pattern is the strong downwarping
of shorelines above S2 toward the lake outlet, along both
the northeast and northwest shores of the lake. As
tentatively correlated, these shorelines show downwarping
equal to 2/3 or more of the total differential uplift from
the caldera rim near Lake Butte to their highest measured
elevations. Segment C in Mary Bay also shows large
anomalous tilting in that shorelines fall as much as 2 m to
the northwest, in an opposite sense to historic tilting.

All shorelines higher than S2 show significant warping
which deviates from the historic uplift pattern. This
suggests the operation of deformation processes which are
more sporadic or episodic than the historic inflation and
may or may not be directly related to magmatic processes.

Mary Bay Area

Some apparently anomalous shoreline warping may be the
expression of longer-term adjustments to inflation and
surface distension. At Mary Bay, the entire sequence of
shorelines above S1 is consistently downwarped to the
northwest, a reversal of the apparent overall trend (Fig.
This anomalous warping takes the form of downfolding, although the lack of preserved shorelines directly to the southeast leaves the net displacements of this feature in question.

It is possible to interpret this deformation in the context of caldera inflation. Although it is possible to resolve only vertical surface displacements by the use of shorelines, the actual mode of surface deformation which occurs with magma-induced inflation is distension. The crust above the expanding magma chamber must extend along its originally horizontal direction in order to accommodate the increased volume of magma. It may be assumed that at shallow crustal levels in the caldera (< 5 km; Smith and Braile, 1984) extension will occur by brittle failure, and that it may occur on favorably oriented pre-existing fractures. This discussion seems at odds with the lack of observed surface rupture in the study area. However, if faulting occurs gradually by stable sliding rather than stick-slip motion, as suggested by Pelton and Smith (1982) as possible for the intracaldera environment, then weak surficial deposits may be folded over the underlying faulted bedrock. This may occur without surface rupture only if offset by normal faulting is small.

Mary Bay is a likely area for fault reactivation to occur. Its location relative to the caldera rim plus the large magnitude of past and present hydrothermal activity
and heat flow (Morgan et al., 1977; Wold et al., 1977) suggests that it lies within the main ring-fracture zone of the caldera (Christiansen, 1984). Evidence of ring-fracture involvement in the present uplift was noted by Pelton and Smith (1982) in the close correspondence of the zero contour of uplift to the outer edge of the ring-fracture zone. Hydrothermal fluids may also aid in fault movement by increasing pore fluid pressures.

The relative offset shown by folding at Mary Bay is down toward the caldera axis, as would be expected for a northwest-dipping ring-fracture. Reactivated faults of the keystone graben on the resurgent dome of the Long Valley caldera show a similar sense of displacement with presently active uplift (Denlinger and Riley, 1984). Although the Long Valley and Yellowstone calderas may not be undergoing resurgence per se, the form of uplift is similar. Reactivated ring-fractures, again with dip-slip motion down towards the caldera center, aided in extension during resurgence of the Timber Mountain caldera in Nevada (Carr and Quinlivan, 1968).

An interesting feature of the warping at Mary Bay is that shoreline S4 is more steeply downwarped than S6, which lies above S4 and is presumably older. This is not necessarily contradictory, since S6 may have been tilted up to the northwest more steeply than S4 prior to downfolding and thus show less net tilting.
Fishing Bridge Area

The strong downwarping of higher shorelines near the lake outlet at Fishing Bridge appears to have no simple relationship to caldera inflation (Fig. 17). It is restricted to shorelines higher than 6 m above datum and appears to be cumulative; that is, the higher, presumably older shorelines show greater downtilting. The highest measured shoreline in the area (at about 26 m above datum), despite large uncertainties in elevations, is not downwarped. It may either lie outside of the area of downwarping (it is 300 m west of the next lowest, downtilted shoreline), or it may have a more complex deformation history than younger shorelines below it. Geometrically, the pattern of downwarping may be explained by either a 1-2 kilometer-wide, roughly north-south trending, troughlike downwarp, or a larger, roughly planar surface that is tilted down to the northwest (such as a tilted fault block).

One interpretation is that the downwarping at the lake outlet is related to movement on a graben structure that has been identified in a reflection profile about 3 km south of Fishing Bridge (Otis et al., 1977) (Fig. 17). The graben faults cut Holocene lake sediments, with a downdropped block of about 1 km width. Although apparently identified in a single profile, the graben is reported to be north-south trending. The downwarping at the lake outlet fits well with
Figure 17. Shoreline elevation data for the Fishing Bridge - lake outlet area projected to an east-west line (shown on map at bottom). Map also shows approximate location of the graben structure described by Otis et al. (1977).
the width, displacement, and reported trend of the graben, and may be the expression of this tectonic feature as it loses displacement to the north. No faults which appear to be the continuation of this trend are mapped either in surficial materials (Richmond, 1977) or in the bedrock of the Sour Creek dome to the north (Christiansen and Blank, 1975).

However, north-south trending normal faults with associated small graben structures cut Holocene shoreline terraces at Eagle Bay, about 20 km south of Fishing Bridge (Fig. 3) (Richmond, 1974). These faults relate to active east-west extension along a north-south trending segment of the Intermountain Seismic Belt (Smith and Braile, 1984). Although seismicity on this trend is reduced within the Yellowstone caldera, many epicenters of 1971-1979 earthquakes plot along the mapped trace of these faults as well as along the projection of these trends to the north (Smith and Braile, 1984). The Fishing Bridge, Lake Hotel, and Stevenson Island areas are particularly seismic, although it is difficult discern a local trend of faulting from epicenter locations.

The Elephant Back fault zone is a N45E-trending system of elongate horsts and grabens that cuts the 150 ka Elephant Back flow, but is not observed to offset Pinedale sediments (Christiansen and Blank, 1975; Richmond, 1977). Its easternmost fault lies a few kilometers west of Fishing
Bridge. It is possible that a northwestward-downtilted fault block on this trend could produce the warping of terraces seen at the lake outlet. Two lines of evidence suggest that this is not the case. Bull Lake-Pinedale interglacial lake silts exposed in the riverbank and roadcut immediately west of the bridge at Fishing Bridge dip uniformly eastward at a few degrees, supporting a trough structure interpretation. This is not conclusive, since the lake sediments are much older than the terraces and may have experienced a varied deformation history; also, it is possible that the dips, although consistent, are due to unrecognized soft sediment deformation. Since the dip of the beds is toward the river, it is unlikely that the dip is due to slump rotation. Also, the trend of the 26 m terrace described earlier does not fit with northwestward downtilting. The downwarping at the lake outlet could not be produced by a troughlike structure on the N45E Elephant Back trend; the shoreline tilts require a more northerly trend. The actual cause of the downwarping remains problematic, and must be resolved by means other than terrace deformation studies.

The location of the lake outlet within the area of downwarping suggests that recurrent subsidence there may have controlled the outlet location. Again, this question is raised but not resolved by the terrace deformation data. It is likely, however, that downwarping of the outlet area
has had some effect on lake level by lowering the relative "spillway" elevation. The magnitude of lake lowering is dependent on the gradient of the outlet stream (i.e., the Yellowstone River) and the distance along the stream through which the downwarping is effective. With a very low outlet stream gradient as exists at present in the first few kilometers below the lake, the effect of purely local downwarping would be insignificant. Given that past stream gradients as well as the northern extent of downwarping are unknown, it is not possible to estimate the control on lake level exerted by downwarping. It is possible that the low present stream gradient at Fishing Bridge is in part due to downwarping.

Controls on Lake Level and Terrace Formation

A question that is central to the understanding to the Holocene history of the Yellowstone Lake basin is: What are the controlling factors and mechanisms for the preservation of lake level stands as discrete terraces? No lakes other than those dammed by glaciers or in closed basins are known to this author to exhibit well-developed, Holocene-age shoreline terraces. Although no lake exists in North America which is closely analogous in geologic setting to Yellowstone Lake, Flathead Lake in northwestern Montana is of similar size and lies in a basin of glacial sediments. While it is certainly large enough to generate wave-cut
shorelines and well-formed terraces, it exhibits no raised shorelines. Evidently, at Yellowstone Lake some unique processes have operated to record a series of individual water level stands through a period of changing water level.

Factors which may produce relative water level changes in a lake basin are essentially of four types: 1) climatic, which change water levels by altering the balance of inflow and outflow; 2) erosional, by lowering the level of the lake outlet; 3) damming effects, such as by glaciers or landslides across outlet streams; and 4) tectonic, by raising or lowering the outlet level and by tilting and warping the lake basin. A combination of these factors may produce the fluctuating water levels necessary for terrace formation. In order for discrete terraces to be formed, it is necessary that either water level change is episodic, with stillstands between periods of change, or that shoreline erosion is episodic, e.g. recording storm events superimposed on a continually changing water level.

At Yellowstone Lake it appears that shoreline erosion, although certainly not constant in rate, is continually occurring. Rates of erosion in the unconsolidated materials making up most of the coast are relatively rapid, with abundant evidence of active undercutting and slumping of wave-cut cliffs visible over much of the shoreline. Richmond (1976a) observed the wave-cut cliff near Squaw Lake (where wave fetch is near maximum for the lake) to be eroded
as much as 1.2 m laterally in less than one hour during a storm. Thus it seems unlikely that storm events are isolated enough in time and unique enough in erosional power to be recorded as terraces. Storm events do account for the preservation of water level markers as beach ridges on marine coasts undergoing slow, relatively constant isostatic uplift (e.g., Blake, 1975; Hillaire-Marcel and Fairbridge, 1978). These coasts experience storm tides and waves which may strand beach gravels several meters above normal water level erosion. The formation of spit and bar crests during storms at Yellowstone Lake is somewhat analogous, but these features do not occur in sets of successively younger, lower ridges as do beach ridges in the Arctic.

The seasonal water level fluctuation at Yellowstone Lake is climatically controlled, involving the balance of precipitation, snowmelt, runoff, infiltration, evaporation, and other factors. In a closed basin, change in the balance of these factors can result in lake level changes of tens to hundreds of of meters, and shoreline terraces may be formed during periods of relative climatic stability. The Pleistocene pluvial lakes of the southwest commonly exhibit terraced shorelines (e.g., Smith and Street-Perrott, 1983). In a lake with an outlet, however, water level change is limited to the change in height of the outlet stream through its range of discharges. As stated earlier, this variation is less than 2 m for Yellowstone Lake at present (Denison et
al., 1982). Since there is no reason to believe that the lake has existed as a closed basin at any time during the Holocene, climatic change can be rejected as a factor producing significant water level changes.

Outlet erosion is a mechanism for lowering of lake level that has without question operated throughout the Holocene. If outlet erosion proceeds gradually and without cessation, however, no stillstands of water level are produced that may be recorded as terraces. Catastrophic outlet erosion can rapidly change lake level, such as at Lake Bonneville (Malde, 1968). The rapid erosion of the Red Rock Pass outlet there, however, was apparently caused by lava-flow damming of the Bear River drainage, causing capture of the drainage by Lake Bonneville and consequent rapid rise of the lake to the outlet level (Bright, 1967). Surficial materials at the pass were rapidly stripped away to resistant bedrock, where the outlet level stabilized. In the absence of any geologic evidence of Holocene flood events in the Yellowstone Lake outlet, it is assumed that stream energy available for erosion has been more or less uniform.

Since discharge at the outlet is relatively constant on an annual maximum basis, intermittent erosion must be imposed by the materials being eroded. It is difficult to understand how this might occur given the stratigraphy of glacial materials at the lake outlet. Glacial sediments
that were silica-cemented by hydrothermal activity may have provided relatively resistant layers to outlet erosion, as suggested by Richmond (1976a); however, these do not commonly occur as thin horizons in the glacial stratigraphy which could produce multiple stillstands. Instead, the coarser sediments are commonly cemented throughout their vertical extent. As at Steamboat Point and Storm Point, this occurs locally around centers of hydrothermal activity, but not as widespread horizons of cementation.

Other variations in outlet material such as between sand and gravel and compact silt might impose significant variations in erosional rates but such variation would not cause intermittent erosion on a scale of hundreds to thousands of years. If the outlet was eroded to a control on bedrock, as it apparently was at LeHardy Rapids, erosion might be slowed enough such that a stillstand would be created. However, it is again difficult to see how this effect alone could occur enough times to form multiple terrace levels.

Damming effects can cause essentially instantaneous water level changes. Ice-damming effects created a series of multiple shorelines around Glacial Lake Missoula (Pardee, 1942), and are also apparently responsible for Late Pleistocene terraces above Yellowstone Lake (Richmond, 1976a). This mechanism would of course not be effective for any Holocene-age terrace. No geologic evidence is seen for
landslides, lava flow dams, or other depositional events which might have blocked the Yellowstone River and raised lake level during the Holocene. Moraine "dams" may have had some effect in retaining the lake in the early Holocene, but it is not plausible that these dams should effectively halt outlet erosion until some breaching event, unless the event consisted of opening and occupation of some entirely new, lower outlet channel. Again, there is no evidence that the Yellowstone River has followed a course significantly different from its present one at any time during the Holocene.

Perhaps the most likely mechanism for quantum lake level changes is magma-related or tectonic vertical surface movements. Abundant evidence for Holocene vertical deformation of the lake basin is seen in the results of this study. The most significant effect of the present uplift on lake level is to raise the outlet more rapidly than any point on the lakeshore, so that unless erosion can fully compensate, relative water level will rise over the entire lake basin. Lake gage records show that this is in fact occurring, with net uplift of the outlet control (uplift minus erosion of the control) at about 12.5 mm/yr (Hamilton, 1985a). As caused by this mechanism, lake level rise relative to a point on the lakeshore is dependent on the concurrent rate of vertical deformation at that point. Evidence for a recent rise in lake level is also seen in
drowned and dying trees lining the Southeast Arm (Hamilton, 1985a), where relative rise of the lake would presumably be greatest given the pattern of present uplift. The main control on lake level is presently on the exhumed bedrock surface at LeHardy Rapids, 5 km downstream from the nominal outlet, with control by the nominal outlet only at low water stages. Continued operation of this mechanism will reduce the gradient of the river above the control to zero, effectively turning it into a narrow arm of the lake. With long-term uplift at rates well exceeding the rate of outlet downcutting, the existing shoreline would be submerged around the entire lake.

Evidence for drowned shorelines that may have been submerged by the above mechanism is seen in sonar profiles of the lake bottom (Bailey, 1984; Hamilton, 1985b). Well-formed terrace-like step and riser morphology was revealed at depths of up to 7 m or more below the lake surface. Beach facies sediments were found on the hypothesized terrace platforms, with possible cliff-derived boulders seen near one terrace-like slope break. Closely-spaced sonar profiles around the north end of the lake allowed tentative correlation of slope breaks and suggested warping of submerged shorelines (Hamilton, 1985b), but further work is necessary to confirm the identity of the submerged features and understand their relationship to onshore terraces. Raised shorelines may plunge below present lake level in the
southeastern study area near the caldera boundary, but submerged shorelines on the northernmost and western shores of the lake are probably not extensions of raised shorelines in the study area.

Geomorphic evidence in the outlet area also suggests that at least one major episode of lake level rise has occurred in the later Holocene. About 1 km downstream of Fishing Bridge a large meander scar of the Yellowstone River is cut into glacial deposits, which are also cut by older shoreline terraces at heights of 12 m and above. The channel of this scar is cut at or below the present level of the river, which floods its downstream end. However, much of the upstream end of the old channel has been filled in by a large spit or pendant bar. This feature prograded almost completely across the mouth of the cove formed by the meander scar. It is difficult to determine whether this feature is of fluvial or lacustrine origin, although the recurved morphology visible in air photos suggests that it is a spit. The ability of longshore transport to construct such a feature after flooding of the meander scar is well illustrated by spits at the present nominal lake outlet. The top of the spit or bar lies at about 6 m above datum, indicating a minimum rise in local water level of about 4.5 m. The spit or bar is in turn terraced at 4.3 m above datum, indicating a subsequent stepwise decline of water to its present level. A large spit is built inside the outlet.
on its west side, with material derived by active cliff erosion and longshore drift from the Bridge Bay - Fishing Bridge shoreline segment.

Additional evidence for a past episode of rising lake level was seen in a pipeline trench dug across the Yellowstone River channel just below the bridge at Fishing Bridge. This trench exposed a channel-like cut in the Pinedale-Bull Lake interglacial lake silts filled with at least 4 m of gravel (W. L. Hamilton, personal communication, 1984). This suggests a river channel cut well below present river level and subsequently filled by aggradation during rising water levels.

Deformation within the caldera is a probable mechanism for quantum water level changes necessary for terrace formation. However, the pattern of terrace deformation suggests that the Holocene deformational history of the northern lake basin and outlet area is more complex than simple inflation episodes of the present form. Also, other important factors must be considered in order to understand terrace formation. For example, since the effect of inflation is to raise relative lake level over the entire basin, some mechanism must operate as well to lower the outlet level such that previously eroded shorelines are not submerged during subsequent inflation episodes. Outlet downcutting may accomplish some or all of the required lowering. It may be that terrace-forming erosion occurs
mainly during episodes of inflation, since rising water will continue to undercut and erode existing wave-cut cliffs. Shorelines would then be stranded or "raised" during periods of tectonic stability, when water level falls below shoreline angle levels by outlet erosion. It is difficult for lowered water level to effectively undercut the wave-cut cliff, as suggested by the location of shoreline angles near mean high water level. The wave-cut platform is an efficient ramp which dissipates wave energy at lower levels (Pethick, 1984).

Since outlet erosion on bedrock occurs slowly (1.4 mm/yr; Hamilton, 1985a) fall of water level may be aided by magmatic deflation and accompanying subsidence of the outlet control. This would serve to isolate terrace levels from wave erosion more rapidly. Although the shoreline elevation data suggest net caldera inflation, lesser deflation events may also have occurred. Since this mechanism would also cause downwarping of the study area lakeshore, relative lake level decline would be small. A more effective mechanism for rapid decline of water level is localized downwarping of the lake outlet control. As stated previously, it is not possible to quantitatively estimate the effect of outlet downwarping on lake level, but it is quite possibly significant.

It is probable that vertical deformation of both a magmatic-related and a tectonic nature has played a dominant
role in the formation of Yellowstone Lake terraces, yet links between specific deformational events and episodes of terrace formation cannot be established at present. The problem is complex, requiring data on the chronology of deformation and erosion for the entire lake/outlet control system.
CONCLUSIONS

Existence and Nature of Shoreline Deformation

Significant surface deformation involving several meters of net vertical movement has occurred in the northern Yellowstone Lake basin in Holocene time, resulting in tilting of raised shorelines of up to 6000 microradians. This is in contrast to the earlier view of Pelton and Smith (1982) which, based on limited available data and assumed correlations of terraces (Richmond, 1976a), supported little or no net Holocene deformation prior to the historic period. The chronology of deformation remains poorly understood, but differences in net deformation between terraces and radiocarbon minimum ages suggest that vertical deformation has been ongoing for at least 2500 years. The discontinuous nature and complex warping of terraces, plus the lack of applicable correlation methods, have prevented positive correlations; therefore shoreline deformation is interpreted here through tilting patterns of individual terrace segments and tentative correlations between them. Using a radiocarbon minimum age of approximately 2500 yr B.P. for a relatively continuous higher terrace, a derived tilting rate is 2/3 or less of historic rates. Assuming that the caldera rim is a stable base at which no Holocene deformation has
occurred, the uplift rate since 2500 yr B.P. near Lake Lodge is 40% or less of historic rates. Rate differences between Holocene and historic uplift as well as cumulative deformation of older shorelines suggest that vertical deformation has been episodic; reversals in direction may have also occurred.

The location of Holocene and historic deformation within the Yellowstone caldera strongly suggests that active magmatic processes are responsible (as discussed by Pelton and Smith, 1982, p. 2755-2758); however, no shoreline elevation data is as yet available from outside of the caldera to determine the boundaries of Holocene deformation. Where it exists, historic leveling data in Yellowstone indicates a relatively simple pattern of inflation of the entire caldera. The pattern of shoreline deformation suggests that inflation of a similar type has occurred, with the additional influence of more varied deformation sources. These may include local magmatic sources such as dike or cupola intrusion, extensional deformation related to uplift, and tectonic regional extensional deformation essentially unrelated to specific magmatic events. Glacio-isostatic uplift may have played some role in deformation of early Holocene shorelines, but the data at hand cannot resolve this question.

No discrete offset of terraces by faulting was noted, but significant local warping suggestive of faulting was
observed at Mary Bay and near the lake outlet at Fishing Bridge. Evidence of surface rupture at these locations may have been removed by later lacustrine and fluvial erosion.

Lake Level Chronology and Controls on Terrace Formation

Existing radiocarbon dates support a roughly constant decline of lake level since ca. 9000 yr B.P., but inaccurate dating and problems of terrace correlation between widely spaced localities prevent detailed understanding of lake level chronology. Various lines of geomorphic evidence including the possible existence of submerged terraces suggest that lake level decline has been interrupted, perhaps several times, by episodes of rising water level. By controlling such water level fluctuations, vertical surface deformation may have played a dominant role in the formation of shoreline terraces. As at present, caldera inflation causes differential uplift of the lake outlet and accompanying water level rise throughout the lake basin. Opposing rates of outlet downcutting are largely dependent on the nature of the material being eroded (e.g., unconsolidated vs. indurated). Lowering of lake level may have also occurred through episodic local downwarping of the outlet area.

In accordance with the sum of the above factors, terraces are probably cut during periods of stable to rising water levels, when wave-cut cliffs may be effectively
undercut. Falling water levels result in damping of wave energy on existing wave-cut platforms and eventual stranding and preservation of shoreline angles. The existence of at least seven discrete Holocene terrace levels thus implies equally numerous episodes of deformation.

Implications for Future Volcanic Activity

Though the Holocene trend of surface deformation at Yellowstone appears to indicate net inflation of the caldera, probable magnitudes of inflation (ca. 10 m maximum uplift) are not large relative to inflation observed at monitored calderas. This suggests that major surface uplift prior to a climactic pyroclastic eruption is not presently occurring, although whether the underlying magma chamber is presently evolving toward such a state cannot be determined. Premonitory seismicity suggestive of ring fracture generation or reactivation is also not presently observed. Because no large caldera-forming pyroclastic eruption has ever been directly observed such statements are necessarily deductive.

Historically, observed eruptions at large calderas have been relatively small lava flow and pyroclastic events generally preceded by rapid inflation (on the order of mm to cm/day) and swarm seismicity. Recurrence intervals and duration of postcaldera volcanism at Yellowstone and other large silicic calderas suggest that flow and pyroclastic
eruptions of the Plateau Rhyolite type are possible at Yellowstone, and are even probable on a scale of $10^4$-to-$10^5$ years. However, these should be preceded by increased surface deformation and characteristic seismicity as observed elsewhere. More detailed study of records of past deformation as well as continued monitoring of present activity is necessary to understand the state of the Yellowstone system at present.

**Suggestions for Further Study**

If the Holocene deformation history of the Yellowstone Lake basin and the relationship between shoreline terraces and deformation is to be understood to the fullest extent possible, then a complete system approach in examining terrace formation is necessary. An important element of the Holocene system not determined in this study is the history of lake level stands. If shorelines outside of the caldera are essentially undeformed, they may provide baseline data on lake level change from which absolute magnitudes of vertical deformation can be determined. Reliable absolute dating of these shorelines and correlation with those inside of the caldera will allow calculation of rates of deformation between and since shoreline formation. The question of glacio-isostatic uplift may also be addressed by examining shorelines little or unaffected by tectonic or magma-related deformation.
An equally important element is deformation at the lake outlet which may have affected lake level. River terraces and lake terraces higher than those measured in this study lie between the nominal outlet and Le Hardy Rapids; these record vertical movements in the outlet control area and should be surveyed in detail.

The work of Bailey (1984) and Hamilton (1985b) suggests that a significant part of the shoreline record lies beneath the surface of Yellowstone Lake. The technical difficulties of correlation and dating of submerged terraces are greater than for raised terraces, but side-scanning sonar equipment may be able to resolve three-dimensional lake bottom morphology rapidly and accurately and greatly enhance correlation ability.

Establishment of reliable correlations between terrace segments remains a primary concern if further progress is to be made. Obsidian hydration dating offers a possible method of correlation for which samples are readily available on virtually all terrace segments. The fracture and abrasion history of obsidian grains in beach sediment must still be examined in order to judge the suitability of the method.

The radiocarbon dating potential of all sediments of the lacustrine shoreline environment should be further explored. Beach facies sediments of various terrace levels are frequently exposed in the modern wave-cut cliff and may contain datable organic materials. Dating of lagoon-fill
Silts may be informative if sufficient organic material is present; however, these sediments may contain a significant component of older carbon inherited from Pleistocene sediments as well as younger rootlet contamination. Although no volcanic ash was identified in lagoonal sediments in this study, 6600 yr B.P. Mazama ash may yet be found in this environment.

Detailed mapping of shorelines would allow accurate spatial relationships of elevation data to be presented, and contouring of individual shoreline deformation once correlations had been established. Mapping will become more practical in the near future with the publication of 1:24,000 scale topographic maps of Yellowstone Park. These will provide superior base maps to the 1:62,500 scale maps available at present.

Much of the deformation history recorded in the Yellowstone Lake shoreline terraces remains to be unravelled. Further work will require considerable manpower and expense. However, the deformation record of the terraces of Yellowstone Lake is unique and can provide a rare and valuable insight into the workings of one of the world's largest active volcanic systems.
REFERENCES CITED


Hamilton, W.L., 1985a, Applying lake level gaging records to the investigation of uplift within the Yellowstone caldera, Yellowstone National Park: manuscript in progress.


