Obsidian hydration dating of naturally worked sediments in the Yellowstone region, Montana and Wyoming

by Kenneth Donald Adams

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Earth Sciences

Montana State University

Abstract:
The purpose of this study was to develop an obsidian hydration dating technique that could be applied to naturally worked sediments. Two field settings were chosen. The first setting is a group of five post-glacial fluvial terraces cut into an obsidian-rich outwash plain near West Yellowstone, MT. The second field setting is a group of five raised and deformed Holocene lakeshores near Breeze Point on Yellowstone Lake, WY. Results from the terraces imply that grains are only partially reworked, so that each grain can potentialIy record several events. The dominant reworking mechanism in the fluvial environment is fracturing as opposed to abrasion. All of the terraces were cut in a shorter period of time than the technique can discern. Using the calibration curve from Pierce and others (1976) and a hydration value of 6.5 ±0.5 microns, an age of 20±3 ka is obtained for the terraces. An alternate method using hydration rate constants for the Obsidian Cliff flow (Michels, 1985), an effective hydration temperature (EHT) of 1.4 °C and a rind value of 6.5 ± 0.5 microns yields an age of 15,520 + 2480 -2300 yrs. Both of these ages are older than those calculated by Nash (1984) for the same terraces. However, if the calibration I^C date of 7,100 ± 50 yrs B.P. used by Nash (1984) is too young, his calculated terrace ages are also too young.

Results from the Holocene lakeshores imply coastal processes do not serve as an effective mechanism to fracture obsidian gravel. All of the shorelines possess a dominant rind signal at about 5.5 ± 0.5 microns. Hydration rate constants for the Obsidian Cliff flow (Michels, 1985) were used instead of constants calculated for the Aster Creek flow (Michels, 1988), which forms the local bedrock in the area, because the latter are thought to be in error. Using Obsidian Cliff constants, an EHT of 1.0 °C and a rind value of 5.5 ± 0.5 microns, the shorelines yield an age of 12,270 +2330 -2130 yrs. Considering that all dated postglacial shorelines in the Yellowstone Lake basin are younger than about 9 ka, it is concluded that the rind signal associated with the shorelines probably dates from latest Pinedale time.

All of the terraces and shorelines have rinds that are thicker than those used to date these features. It is hypothesized that these older peaks may date from reworking by Eowisconsin glaciers, Bull Lake glaciers and hydration dating from original cooling cracks, in order of increasing thickness and age.

Hydration rate is dependent on temperature and the chemistry of the glass. Because chemistry varies within and between flows, rate would also be expected to vary. Much of the observed spread in rind thicknesses is probably due to natural variation in hydration rate. Therefore, it is not appropriate to use a single set of hydration rate constants determined from a single sample to calculate ages for artifacts and geological samples without recognizing this limitation.

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APPROVAL

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Kenneth Donald Adams

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.

3/28/90
Date

W W Locke III
Chairperson, Graduate Committee

Approved for the Major Department

3/28/90
Date

Stephen G Custer
Head, Major Department

Approved for the College of Graduate Studies

April 4, 1990
Date

Henry L Parsons
Graduate Dean

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ABSTRACT

The purpose of this study was to develop an obsidian hydration dating technique that could be applied to naturally worked sediments. Two field settings were chosen. The first setting is a group of five post-glacial fluvial terraces cut into an obsidian-rich outwash plain near West Yellowstone, MT. The second field setting is a group of five raised and deformed Holocene lakeshores near Breeze Point on Yellowstone Lake, WY. Results from the terraces imply that grains are only partially reworked, so that each grain can potentially record several events. The dominant reworking mechanism in the fluvial environment is fracturing as opposed to abrasion. All of the terraces were cut in a shorter period of time than the technique can discern. Using the calibration curve from Pierce and others (1976) and a hydration value of 6.5 $\pm$ 0.5 microns, an age of 20 $\pm$ 3 ka is obtained for the terraces. An alternate method using hydration rate constants for the Obsidian Cliff flow (Michels, 1985), an effective hydration temperature (EHT) of 1.4 °C and a rind value of 6.5 $\pm$ 0.5 microns yields an age of 15,520 + 2480 -2300 yrs. Both of these ages are older than those calculated by Nash (1984) for the same terraces. However, if the calibration ^{14}C date of 7,100 $\pm$ 50 yrs B.P. used by Nash (1984) is too young, his calculated terrace ages are also too young.

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INTRODUCTION

The Problem

Numerous methods have been developed and used to date Quaternary deposits and events, including ^{14}C dating (Terasmae, 1984), dendrochronology (Parker and others, 1984), soil development (Birkeland, 1984), amino acid racemization (Wehmiller, 1984), hornblende etching (Locke, 1979; Hall and Heiny, 1983) and obsidian hydration dating (Pierce and others, 1976; Friedman and Obradovich, 1981). All of these methods attempt to measure a certain property or condition that changes progressively with time. However, given the variety and complexity of field situations faced by Quaternary researchers, few methods can be applied in any given location. Therefore, dating techniques chosen by the researcher are dictated largely by field constraints.

In the Yellowstone region there are abundant rhyolitic obsidian lava flows (Christiansen and Blank, 1972) which provide many opportunities to utilize the obsidian hydration dating technique for Quaternary studies. For example, Pierce and others (1976) used obsidian hydration dating techniques to date glacial events near West Yellowstone, MT. In this region, and in other areas where obsidian is found, obsidian hydration has also been used to date volcanic events (Friedman, 1968; Friedman and Obradovich, 1981) and fluvial terraces (Lundstrom, 1986) in addition to the more common usage for archeological applications (Friedman and Smith, 1960; Clark, 1964; Meighan, 1983). This technique has the advantage of being low cost and relatively straightforward to perform. However, the technique is limited to the extent that there must be obsidian present in sufficient quantities and in appropriate field situations. The purpose of this study is to develop a methodology that can be used to date Quaternary landforms that contain naturally worked obsidian.

This study focuses on obsidian naturally reworked in two depositional settings, the first of which is a flight of fluvial terraces near West Yellowstone, MT. These terraces were formed by progressive downcutting of the Madison River through an obsidian-rich outwash plain in post-Pinedale time (Nash, 1984). This is an ideal site at which to apply obsidian hydration dating to fluvial terraces because the same terraces were dated by ^{14}C and the scarp degradation method (Nash, 1984), thus allowing an independent check against the ages obtained in this study.

The second setting is a group of raised and deformed shorelines near Breeze Point at Yellowstone Lake, WY. These shorelines are also post-Pinedale in age and have probably been raised relative to lake level and deformed by a complex interaction of volcano-tectonics and downcutting at the outlet of Yellowstone Lake (Meyer and Locke, 1986). The successful development of a methodology to date these shorelines should provide a means to determine recent rates of deformation in the Yellowstone Caldera.

The fluvial terraces were deliberately chosen first because such surfaces provide an unambiguous event stratigraphy with the uppermost terrace being the oldest (Thompson and Jones, 1986). This assumption allowed for the development of the technique and interpretation of the results without taking into account the possibility that the highest surface may not be the oldest.

Contemporary uplift data reported by Dzurisin and Yamashita (1987) for the Yellowstone Lake area indicate that some areas of uplift have reversed to subsidence. In addition, Baily (1984) and Hamilton (1985) have documented submerged shorelines which also indicate subsidence. Both of these lines of evidence lead to the conclusion that the shorelines around Yellowstone Lake cannot be assumed to be in a relative sequence with the uppermost shoreline being the oldest.

Previous Work

Ross and Smith (1955) first noted that surfaces on weathered obsidian had undergone a chemical and physical change known as hydration. Hydration is the intake of water, which causes an increase in density and refractive index in addition to causing the hydrated layer to become mechanically strained (Ross and Smith, 1955). The difference between the refractive indices of the hydrated and nonhydrated portions of the obsidian and the increased birefringence due to mechanical strain make the hydrated layer easily visible with a standard petrographic microscope (Friedman and Smith, 1960).

Further study by Friedman and Smith (1960) determined that hydration occurs under normal atmospheric conditions and that rate is largely dependent upon temperature and chemical composition of the glass. Therefore, it is necessary to know both the temperature at which hydration occurred and the composition of the glass in order to accurately determine time elapsed since hydration began. The composition of a given sample can be determined by standard analytical techniques, but the effective hydration temperature must be estimated if hydration has spanned a considerable amount of time, given the evidence for Quaternary climatic change (Shackleton and Opdyke, 1973).

Friedman and Long (1976) experimented with several types of obsidian held at various temperatures for different lengths of time to determine the Arrhenius equation relating hydration rate to temperature. This equation takes the following form:

$$k = Ae^{-E/RT}$$

where k is the hydration rate (microns squared per 10^3 years), A is a constant, E is the activation energy of the hydration process (calories per mole), R is the gas constant (calories per degree per mole), and T is absolute temperature (Kelvins). Rate is expressed as microns squared per 10^3 years because hydration plotted against temperature follows a logarithmic curve. Friedman and Long (1976) also determined the relationships between silica content, refractive index and hydration rate. Thus, if the temperature to which an obsidian sample has been exposed and its chemical composition or refractive index are known, the rate of hydration can be calculated. Friedman and Long (1976) concluded that if 1) the effective hydration temperature of a sample is determined by direct measurement or by estimation through climatic records and 2) the hydration rate is determined empirically or calculated from silica content, refractive index, or chemical index then numerical dating to $\pm$ 10 % accuracy is possible. This principle has been applied to naturally reworked obsidian in several studies in the Yellowstone region.

Pierce and others (1976) related the thickness of hydration rinds on pebbles in till to glacial events near West Yellowstone, MT. The measured rinds fell into five distinct groups with the thinnest (youngest) rinds associated with Pinedale deglacial deposits and the thicker (older) rinds with Pinedale terminal moraines, cooling cracks on the West Yellowstone flow, Bull Lake terminal moraines and cooling cracks on the Obsidian Cliff flow, in order of increasing thickness. The two groups of rinds associated with original cooling cracks were calibrated against K-Ar ages for the West Yellowstone and Obsidian Cliff flows to yield approximate ages for the glacial events by interpolation.

Friedman and Obradovich (1981) dated volcanic events at several localities in the western U.S. using obsidian hydration dating techniques. The ages for these events range from 12 ka to over 1 Ma and the hydration dates were compared with ages obtained through ^{14}C and K-Ar methods. In most cases the agreement between the two ages was "good" ($\pm$ 20 %), thus establishing the reliability of the obsidian hydration dating method for volcanic events.

More specifically, Friedman and Obradovich (1981) conducted part of their research in Yellowstone Park, WY. All of the obsidian used for hydration dating was similar in chemical composition to the Obsidian Cliff samples used for experimental determination of hydration rate. A temperature of -1° C was used for the glacial temperature because most of the sites were covered by a large temperate ice sheet (Pierce, 1979). Present day temperatures were determined using the Pallmann technique of temperature integration (Friedman and Norman, 1981). Hydration rind ages were plotted with $\pm$ 20 % uncertainty of the age while the K-Ar ages were plotted with their published uncertainties. As in other regions of the plotted with their published uncertainties. As in other regions of the western U.S. where Friedman and Obradovich (1981) conducted their research, the agreement between the ages obtained through K-Ar dating and obsidian hydration dating was good.

Lundstrom (1986) related the thickness of hydration rinds to a group of fluvial terraces along the Madison River below Quake Lake in southwest Montana. The methodology was not clearly stated nor were the data listed from which the results were derived. Therefore, the interpretation postulated by Lundstrom (1986) cannot be objectively evaluated.

Quaternary Geology of Field Sites

West Yellowstone Basin

The late Cenozoic, extensional, West Yellowstone Basin contains abundant evidence for at least two major glaciations in the Pleistocene that have been correlated with the type sections of the Bull Lake and Pinedale glaciations in the Wind River Mountains of Wyoming (Alden, 1953; Richmond, 1964; Witkind, 1969; Pierce and others, 1976; Pierce, 1979). The floor of the basin is covered by up to 30 meters of obsidian-rich gravel that lies between the more extensive Bull Lake moraines and the smaller Pinedale moraines (Fig. 1) (Richmond, 1964). The age and origin of this gravel deposit has been a matter of debate among the several authors who have worked in this area.

Alden (1953) described the gravel deposit as a large terrace postdating the Bull Lake moraines at Horse Butte. This interpretation was based on the observation that the deposit does not appear to have

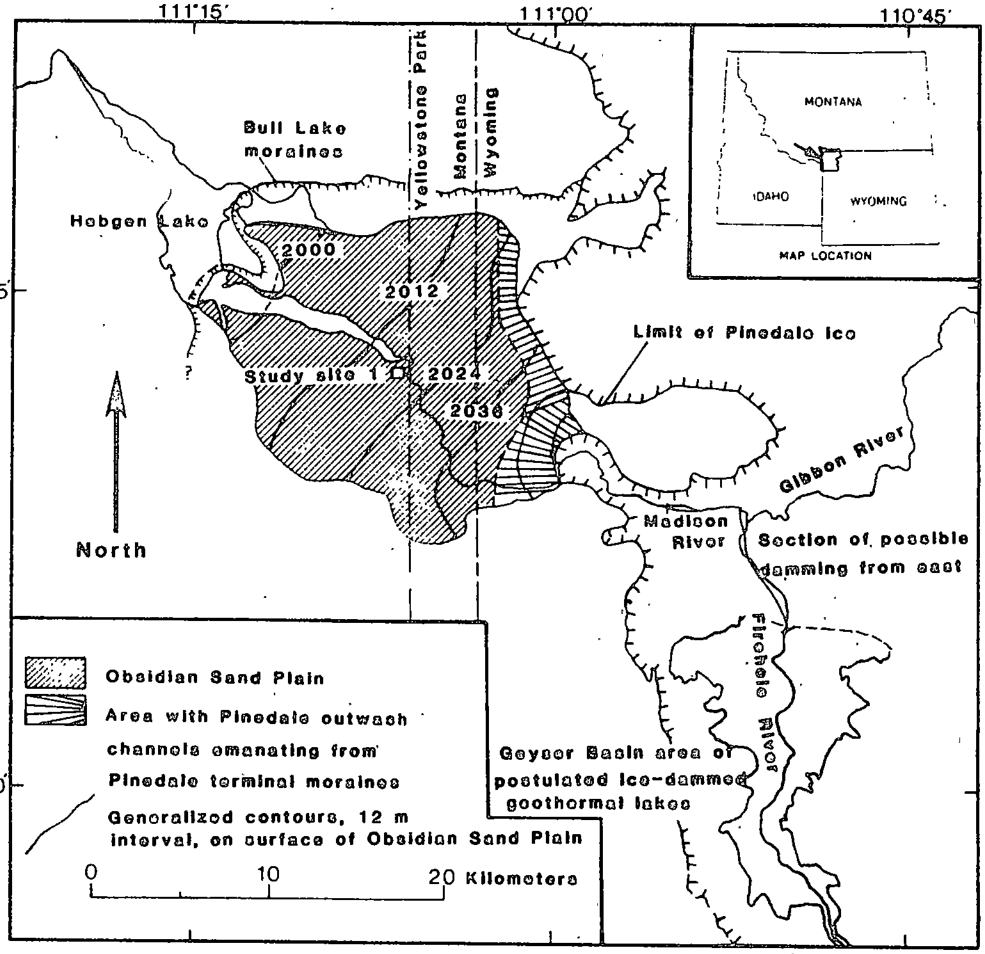

Figure 1 - Map showing study area in relation to Pinedale ice margin. Major elements of glacial outburst flood hypothesis are also shown (after Pierce, 1979).

been overridden by a glacier. Alden (1953) hypothesized that the Wisconsin-age moraines of Beaver Creek, downstream from Hebgen Lake, may have dammed the narrow Madison Canyon, thereby creating a large lake in the West Yellowstone Basin into which the obsidian gravel was deposited.

Richmond (1964) reported that the Obsidian Sand Plain (OSP) was deposited in the Bull Lake-Pinedale interglaciation. This interpretation was based on similar soil characteristics developed on a loess cap found on both the sand plain and the Bull Lake moraines at Horse Butte. Thus the sand plain was deposited after recession of Bull Lake ice but before the soil forming period of the Bull lake-Pinedale interglaciation (Richmond, 1964). He also stated that the structure, texture and especially the lithology of the obsidian gravel suggest that it is unrelated to glaciation.

Richmond (1964) suggested that because the gravel occurs in beds 10 to 30 cm thick and displays crossbedding and channel-and-fill structures, it was deposited by shallow aggrading streams. Much of this sediment may have been derived directly from the surfaces of obsidian flows within the park which contain abundant sand and gravel size material, or alternatively could also have been derived through the fluvial reworking of larger clasts (Richmond, 1964).

Witkind (1969) described the OSP as having been deposited by meltwater from receding Bull Lake ice. Evidence for this is the stratigraphic relationships at Horse Butte and along the north edge of the basin where the gravel overlies till of Bull Lake age.

Pierce and others (1976) suggested that the OSP is Pinedale-age outwash. Evidence includes hydration rinds from one locality on the deposit, which do not differ significantly in thickness from those in Pinedale moraines (Pierce and others, 1976). It must be emphasized that this evidence does not rule out the hypothesis of a composite age for the OSP.

Pierce (1979) maintained that the OSP is Pinedale-age outwash and further postulated multiple glacial outburst floods for the origin of the sand plain (Figure 1). Geothermal heat could have formed a lake 100 km^2 in area and 100 m deep in the Upper, Midway and Lower Geyser Basins (Pierce, 1979). Release of this lake by glacial dam failure at National Park Mountain would have released catastrophic floods into the West Yellowstone Basin. Ponding of the flood waters by moraine damming at Beaver Creek, 1.5 km west of Hebgen dam, may have caused a low gradient flood delta to accumulate in the Basin (Pierce, 1979).

In several areas, the gravel deposit consists of sheetlike beds 15 to 60 cm thick with planar cross stratification inclined at both low and high angles to the bedding (Pierce, 1979). Silt-rich interbeds between crossbedded layers are common and cut-and-fill structures are poorly developed. The crossbeds of fine gravel generally have open boxwork structure and very little interstitial fines. These sedimentary characteristics suggest rapid deposition by sheet flood (Pierce, 1979).

The strongest evidence that Pierce (1979) presented for a Pinedale age for the Obsidian Sand Plain is the documentation of a series of outwash channels emanating from a Pinedale terminal moraine complex at the mouth of Madison Canyon onto the OSP. Thus, the moraines are physically tied to the sand plain.

Pierce (1979) also stated that hydration rinds from the OSP collected at depths from two to twelve meters are the same thickness as those from the Pinedale terminal moraines. However, this data was neither presented nor cited, so it is not known whether it was the same data from Pierce and others (1976) or resulted from additional studies.

Pierce (1979) disagreed with Richmond (1964) in assigning the soils on the OSP to the Bull Lake-Pinedale interglacial. Instead, Pierce (1979) described the soils as weak and similar to profiles developed on other Pinedale-age sandy deposits. The silt in the upper 50 cm of the profiles is attributed to eolian influx and subsequent translocation.

From the above discussion, it is evident that the age and origin of the OSP is a matter of quite some debate. It is likely that the most recent period of deposition occurred during the Pinedale glaciation, but it is also possible that earlier periods of deposition contributed to the bulk of the deposit.

The OSP was dissected by the Madison River in post-Pinedale time leaving five well developed, north facing, fluvial terraces. Pierce and others (1976) reported that the OSP has nearly identical rind thicknesses as Pinedale terminal moraines dated from 28,000 to 40,000 yrs. Because the terraces are inset into this deposit, they must be younger than this age range. Nash (1984) dated the terraces using ^{14}C and scarp degradation methods and assigned ages of 11,600 $\pm$ 6,400, 9,600 $\pm$ 5,600 and 7,100 $\pm$ 50 years to Terraces T2 through T4, respectively. These ages serve as an independent check against the dates for the terraces obtained in this study.

Part one of this study devises an obsidian hydration dating method that can be applied to fluvial deposits to obtain numerical dates. The three main hypotheses that will be tested are: 1) the OSP was deposited solely in Pinedale time and that the hydration rinds from the deposit reflect this age, 2) fluvial processes of traction transport effectively rework obsidian pebbles so that each terrace possesses an unique, characteristic hydration rind thickness, and 3) the dates derived from this study are directly comparable to those obtained by Nash (1984) for the same terraces.

Yellowstone Lake Basin

Yellowstone Lake occupies an irregularly shaped basin approximately 30 km by 25 km that was formed as a result of the complex interaction between caldera collapse and postcollapse volcanism, ice-cap glaciation, and lacustrine processes. The Quaternary history of the basin involves the sequence of these events.

Three major caldera-forming eruptive cycles occurred in the Yellowstone region in late Pliocene and Quaternary time (Christiansen, 1984). Each of these cycles culminated in a major eruption of explosively ejected material that formed voluminous ash flows, but was characterized at the beginning and end by long periods of intermittent rhyolitic and basaltic eruptions. The first cycle began about 2.2 Ma with small eruptions of rhyolite and basalt and culminated with the eruption of the Huckleberry Ridge Tuff at 2.0 Ma. This eruption was the largest of the three caldera-forming eruptions and ejected a volume of material greater than 2500 km^3. Christiansen (1984) stated that the Huckleberry Ridge Tuff formed a single cooling unit and was probably erupted in hours or days and not decades or centuries. The caldera formed by this eruption, which is mostly obscured by more recent lava flows, measured approximately 100 km by 60 km.

The second caldera forming eruption occurred at about 1.3 Ma and was the smallest of the three. The caldera formed by the eruption of the Mesa Falls Tuff is located in the Island Park area and is wholly contained within the Huckleberry Ridge Caldera (Christiansen, 1984). The volume of material ejected from the Mesa Falls Caldera was at least 280 km^3.

The third volcanic cycle began at about 1.2 Ma and, as the most recent cycle, left the most complete record. For the first 600 ka rhyolitic lava was intermittently erupted from a set of arcuate fractures that would eventually outline the Lava Creek Caldera (Christiansen, 1984). At 630 ka, the climactic eruption of the third cycle caldera occurred emplacing more than 1000 km^3 of Lava Creek Tuff. This caldera is about 75 km long and 45 km across. In the last 630 ka, the caldera has been filling with sediments and rhyolitic lava flows, many of which have been erupted in the last 150 ka (Fig. 2) (Christiansen, 1984).

The present shape of the Yellowstone Lake Basin is largely the result of post-collapse volcanism. Christiansen and Blank (1972) proposed a subcaldera collapse associated with the eruption of the Shoshone Lake Tuff at about 162 ka (later informally renamed the tuff of Bluff Point in the West Thumb area; Christiansen, 1974) for the formation of the West Thumb embayment. The remainder of the western shore of

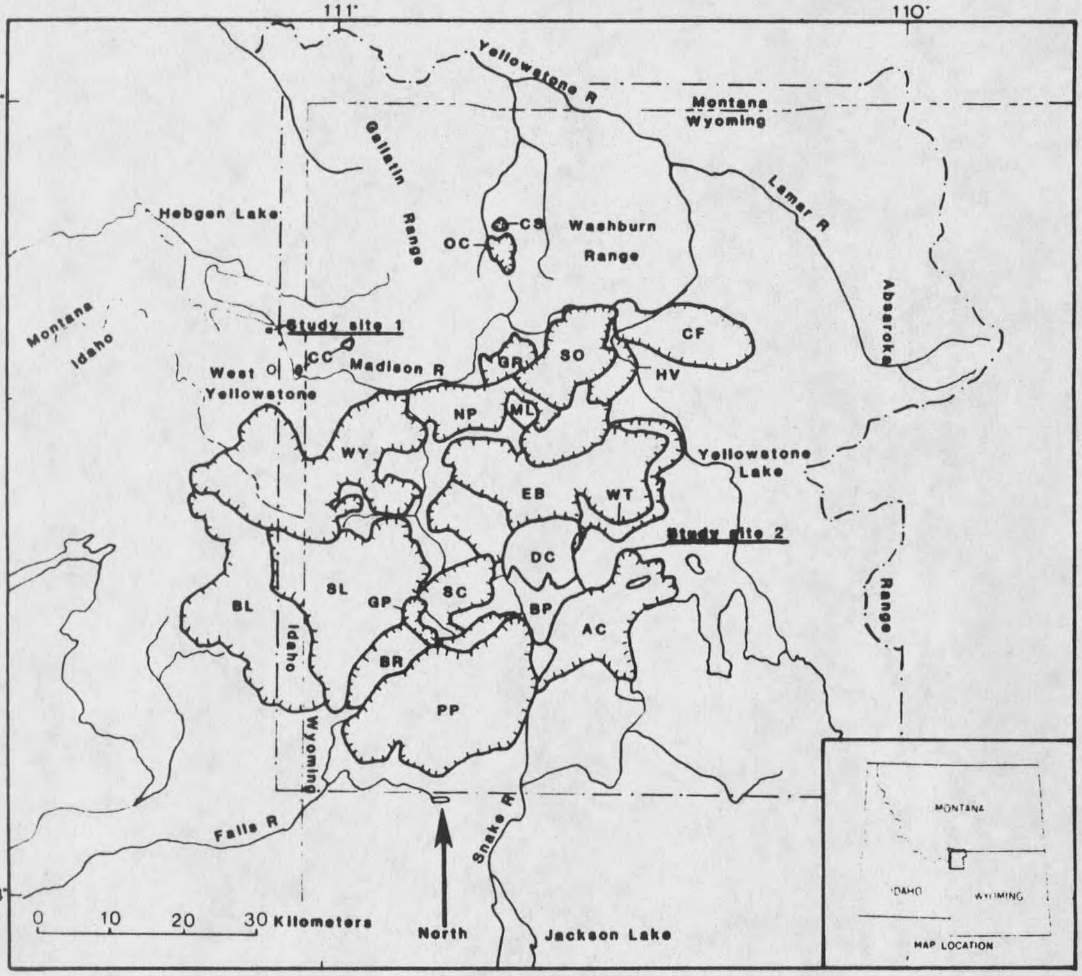

Figure 2 - Regional map of Yellowstone National Park and vicinity showing rhyolite lava flows and location of study sites. Ages of flows (10^3 years) in parentheses (After Richmond, 1986).

AC - Aster Creek flow (174)
BL - Buffalo Lake flow (160)
BP - Tuff of Bluff Point (162)
BR - Bechler River flow (117)
CC - Cougar Creek flow (399)
CF - Canyon flow (613)
CS - Crystal Springs flow (80)
DC - Dry Creek flow (162)
EB - Elephant Back flow (153)
GP - Grants Pass flow (77)
GR - Gibbon River flow (90)
HV - Hayden valley flow (102)
ML - Mary Lake flow (165)
NP - Nez Perce flow (160)
OC - Obsidian Cliff flow (183)
PP - Pitchstone Plateau flow (70)
SC - Spring Creek flow (?)
SL - Summit Lake flow (113)
SO - Solfatara Plateau flow (110)
WT - West Thumb flow (148)
WY - West Yellowstone flow (117)

Yellowstone Lake is bordered by rhyolitic lava flows that probably contributed to the impoundment of the lake (Fig. 2) (Richmond, 1976).

The West Thumb flow erupted eastward across the caldera basin and dammed the Southeast Arm paleocanyon at about 148 ± 12 ka by impinging on the flank of Sour Creek Dome a few kilometers north of Fishing Bridge, causing the formation of an ancestral Yellowstone Lake (Richmond, 1976; Richmond, 1986). The Elephant Back flow K-Ar dated at 153 ± 2 ka overlies the slightly older West Thumb flow and is cut by a series of northeast-trending faults that form a complex graben structure (USGS, 1972).

Other rhyolitic lava flows of the third volcanic cycle that may have played a role in the formation of the Yellowstone Lake Basin and were likely to have contributed sediment to the basin are the Dry Creek flow, K-Ar dated at 162 ± 2 ka, and the Aster Creek flow K-Ar dated at 174 ± 4 ka and which forms the local bedrock in the Breeze Point area (Fig. 2) (Richmond, 1986).

The Yellowstone Plateau, on which Yellowstone Lake is located, has served as the locus for at least three ice cap glaciations and possibly as many as twelve (Richmond, 1986). Glaciation probably began in late Pliocene time but the oldest, well-dated glacial deposits date from the early Pleistocene. Glaciers dating from the Illinoian and Wisconsin periods are known to have filled and overflowed the caldera formed by the eruption of the Lava Creek Tuff. Glaciers that accumulated at about 150 to 140 ka have traditionally been correlated to the Bull Lake Glaciation while glacial advances occurring from 117 to about 80 ka have been reassigned to the Eowisconsin period (Richmond, 1986). Ice that accumulated from >45 to about 10 ka has been correlated to the Pinedale Glaciation (Pierce and others, 1976). The designations of Bull Lake and Pinedale Glaciations follow the terminology of Blackwelder (1915) for the Wind River Mountains of Wyoming (Pierce and others, 1976; Richmond, 1976; Pierce, 1979) whereas glaciers dating from the Eowisconsin follow the terminology of Richmond (1986).

The formation of the present Yellowstone Lake began with the downwasting of Pinedale ice in the lake basin at about 14.5 ka when ice marginal lakes formed around the stagnant ice (Richmond, 1976a). Richmond (1976a) described broad, discontinuous terraces at up to 143 m above the present Yellowstone Lake and concluded that all surfaces above 33 m were kame terraces. A shoreline at 33 m above present lake level shows both ice-contact and wave-cut features which might be expected to form in association with an ice-marginal lake. Surfaces at 27 and 24-23 m are generally discontinuous and poorly formed but are believed to have been formed during minor pauses in the erosion of the Yellowstone Lake outlet (Richmond, 1976a).

Richmond (1974, 1977) mapped post-glacial shorelines at 18-20, 17, 12-14, 9-10.6, 7.6, 4.5, and 3 m above present lake level. Deltaic sands graded to an 18-20 m shoreline at the south end of the Southeast Arm overlie lake silts containing organic matter radiocarbon dated at 9060 $\pm$ 300 yr B.P. (Richmond, 1974). Richmond (1976a) considered this shoreline to be the first to form after complete deglaciation of the lake basin because there are numerous large spits, bars and lagoonal depressions associated with this shoreline that are not noted at higher levels. Such features are indicative of strong longshore currents that are usually associated with large bodies of open water.

Richmond (1974) obtained a minimum radiocarbon date of 5590 ± 250 yr B.P. from a lagoonal deposit for a 12 m shoreline at the south end of the Southeast Arm (Baker, 1976). Organic-rich clay overlying deposits of the Squaw Lake hydrothermal explosion have been radiocarbon dated at 3500 ± 250 yr B.P. The 7.6 m shoreline is cut into the explosion deposit in a bluff of Yellowstone Lake southeast of Squaw Lake. Charcoal found between beach gravel and overlying colluvium on a 3 m shoreline was radiocarbon dated at 620 ± 250 yr B.P. thereby providing a minimum age for this surface (Richmond, 1976a).

Meyer (1986) obtained ^{14}C dates from lagoonal deposits for two shorelines in the northwest part of the lake basin. An age of 2495 ± 135 yrs B.P. was determined for Shoreline S6 (Shoreline S5-this study) at about 18 m above lake datum and an age of 1410 ± 160 yrs B.P. was assigned to Shoreline S5 (Shoreline S4-this study) at about 8 m above lake datum. The dates were derived from basal organic matter that accumulated since the abandonment of the shoreline and may have been contaminated by rootlet intrusion. Therefore, these dates may be several thousand years too young and must be considered minimum dates (Meyer, 1986). Locke (1989) reported, in a poster, two additional minimum ^{14}C dates of about 6 ka for Shoreline S6 (Shoreline S5-this study).

From the above discussion, it can be concluded that the elevated shorelines around Yellowstone Lake were formed during the last 9 ka as indicated by minimum and maximum radiocarbon dates. Richmond (1976a) assumed that the shorelines were horizontal and correlated them by height above present lake level. Further work by Meyer (1986), Meyer and

Locke (1986) and Locke (1989) showed that the raised lakeshores are significantly deformed. This was done by precise leveling surveys and correlation of shorelines between leveling lines. Estimation of vertical deformation range up to 4 mm/yr for shoreline S6 on the northeast shore, compared to the historic rate of 7.5 mm/yr for that area (Meyer, 1986; Pelton and Smith, 1982). Shoreline elevation data from Meyer (1986) also indicate that some shorelines are tilted by as much as 6 m/km. Given the evidence for vertical deformation, reevaluation of shoreline correlations reported by Richmond (1976) is warranted.

The radiocarbon dates given above indicate that lake level has been declining at a roughly constant rate for about the past 9 ka. However, various lines of geomorphic evidence indicate that lake level decline has been interrupted, perhaps several times, by episodes of rising lake level. Bailey (1984) and Hamilton (1985) have documented submerged shorelines which indicate that water level has risen to the present level after forming the now submerged shorelines. Meyer (1986) suggested that caldera inflation as documented by Pelton and Smith (1982) would cause differential uplift of the lake outlet and accompanying water level rise throughout the lake basin. Opposing rates of downcutting at the lake outlet would depend largely on the relative erodibility of the outlet material. Lowering lake level may also be controlled by local downwarping as indicated by shorelines tilting down towards the lake outlet (Meyer, 1986).

Considering the above factors, lakeshores are probably formed during periods of stable to rising water levels, when wave-cut cliffs can be effectively undercut (Meyer, 1986). Periods of rising water level have been roughly equated to episodes of uplift at the outlet. During periods of falling water level the wave energy is presumably dissipated on the wave-cut platform. The existence of at least ten discrete Holocene shorelines implies at least as many episodes of deformation.

The studies of Locke and Meyer (1985), Meyer (1986), Meyer and Locke (1986) and Locke (1989, personal comm.) have done much to decipher the complex volcano-tectonic deformational signal that is displayed in the Yellowstone lakeshores. However, the studies have been restricted by the lack of a means to numerically date the shoreline segments and make firm correlations between widely spaced locations. Meyer (1986) suggested the use of obsidian hydration dating techniques to solve the dating and correlation problem because of the abundance of obsidian gravel on the raised shorelines.

The purpose of the second part of this study therefore, is to devise a method, using obsidian hydration dating techniques, to numerically date and correlate discrete shoreline segments. The hypothesis that will be tested in this part of the study is that lacustrine processes of storm swash and longshore currents effectively rework obsidian pebbles so that each shoreline possesses a unique, characteristic hydration rind thickness.

METHODS

Sampling

Obsidian Sand Plain

Sampling pit locations for study site 1 were selected from a sketch map of fluvial terraces along the Madison River by Nash (1984) (Fig. 3). Sampling pits were located between dated terrace scarps so that age dating results from this study should be directly comparable to those of Nash (1984). This is because terraces and terrace scarps are simultaneously acted upon by the stream and the abandonment of a particular stream position is reflected in the ages of both these types of surfaces.

Sampling pits were dug by the author on each of the fluvial terraces to a depth of one meter and samples collected at 20 cm increments. There was no visible stratigraphy and very little soil development. Soil profiles generally consisted of a thin (5 cm) A horizon with a thicker (20 cm) Bw horizon. Care was taken to dig the pits away from the toe of the adjacent slope so that there was little chance of mixing material from the next higher terrace. Modern river gravel samples were collected from the active river channel and from an exposed gravel bar. The modern gravel samples were collected at a depth of about 20 cm because of the difficulty of digging a pit in the active channel and the proximity of the water table to the surface of the gravel bar.

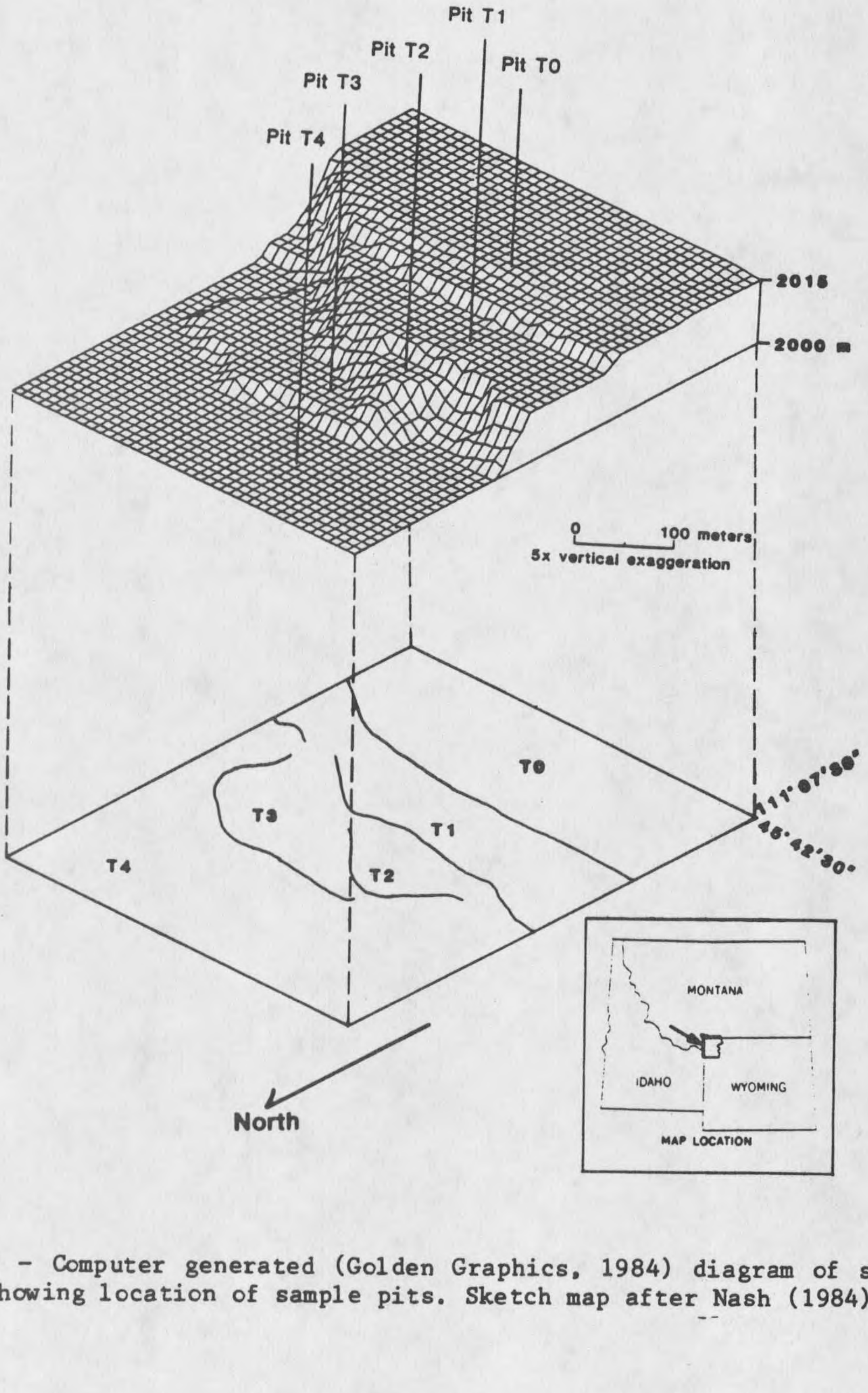

Figure 3 - Computer generated (Golden Graphics, 1984) diagram of study site 1 showing location of sample pits. Sketch map after Nash (1984).