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.

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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 ± 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 ^{14}C date of 7,100 ± 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

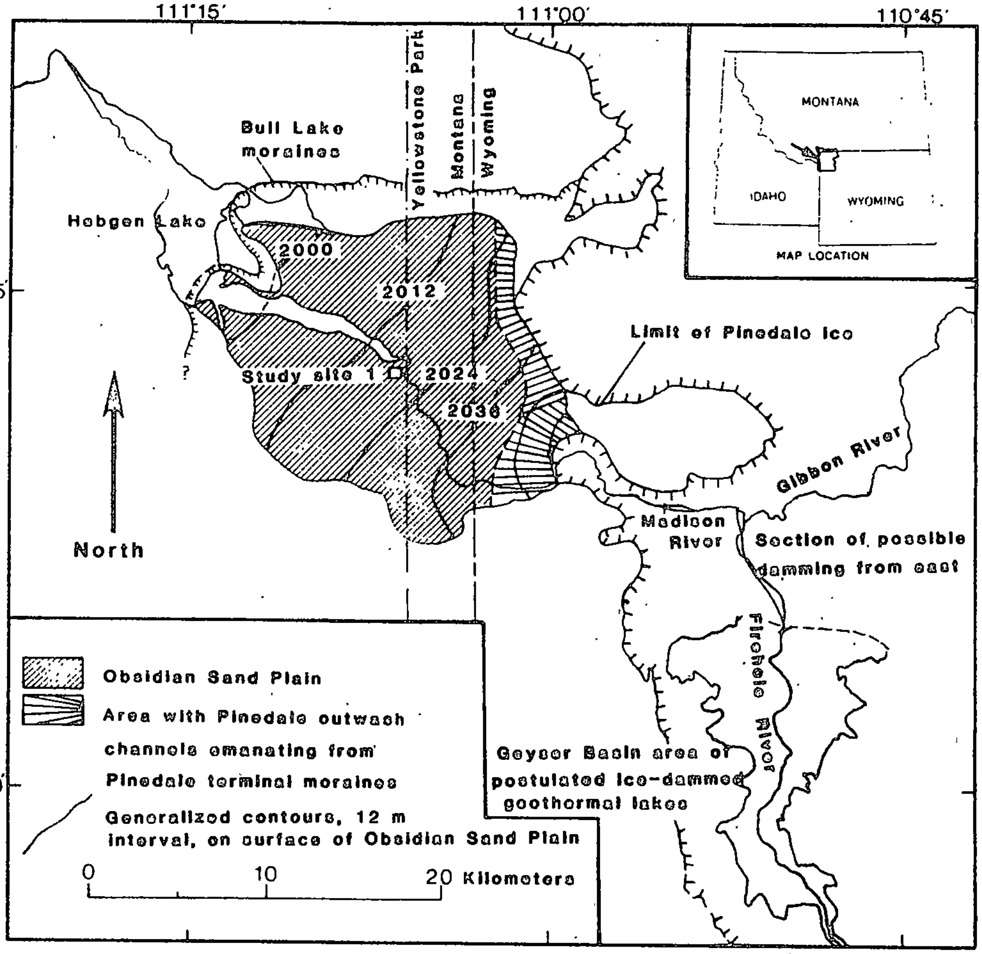


Figure 1 - Map showing study area in relation to Pinedale ice margin. Major elements of glacial outburst flood hypothesis are also shown (after Pierce, 1979).

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

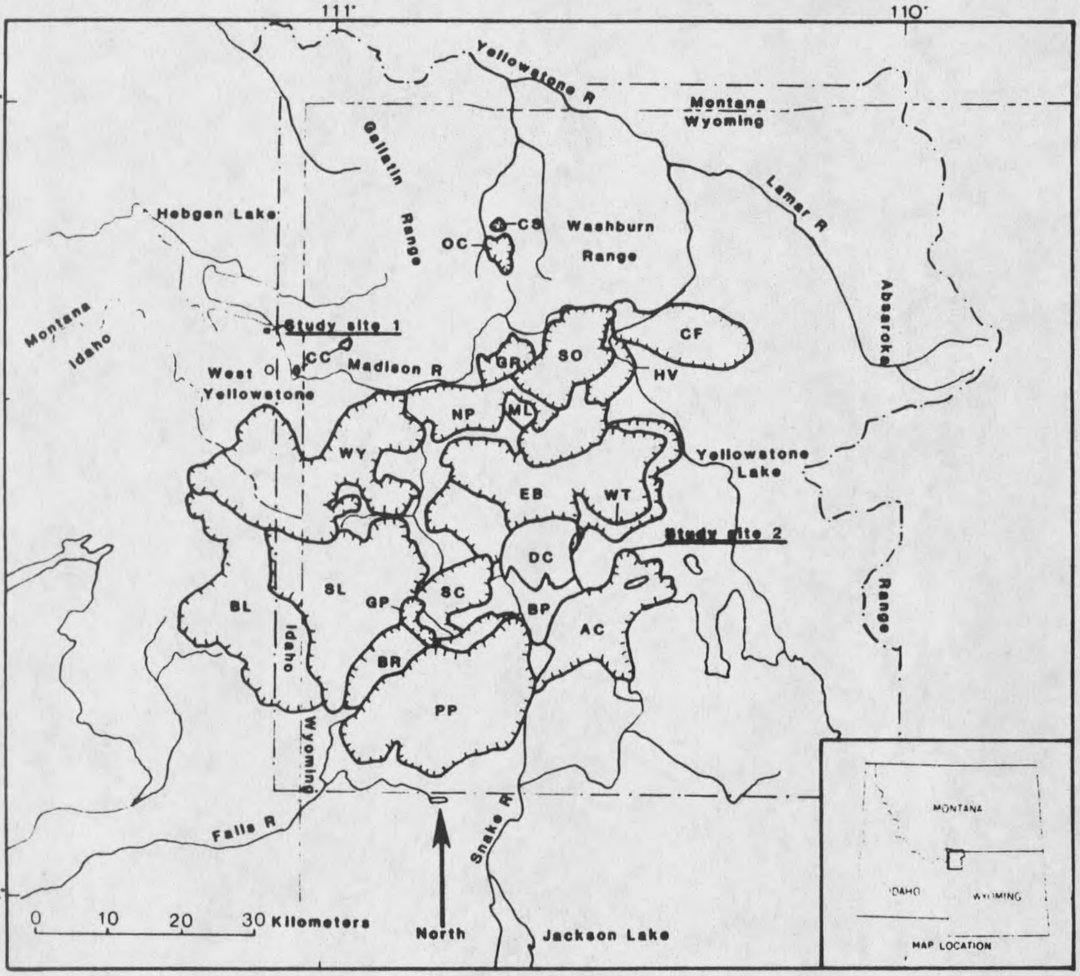


Figure 2 - Regional map of Yellowstone National Park and vicinity showing rhyolite lava flows and location of study sites. Ages of flows (10^3 years) in parentheses (After Richmond, 1986).

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.

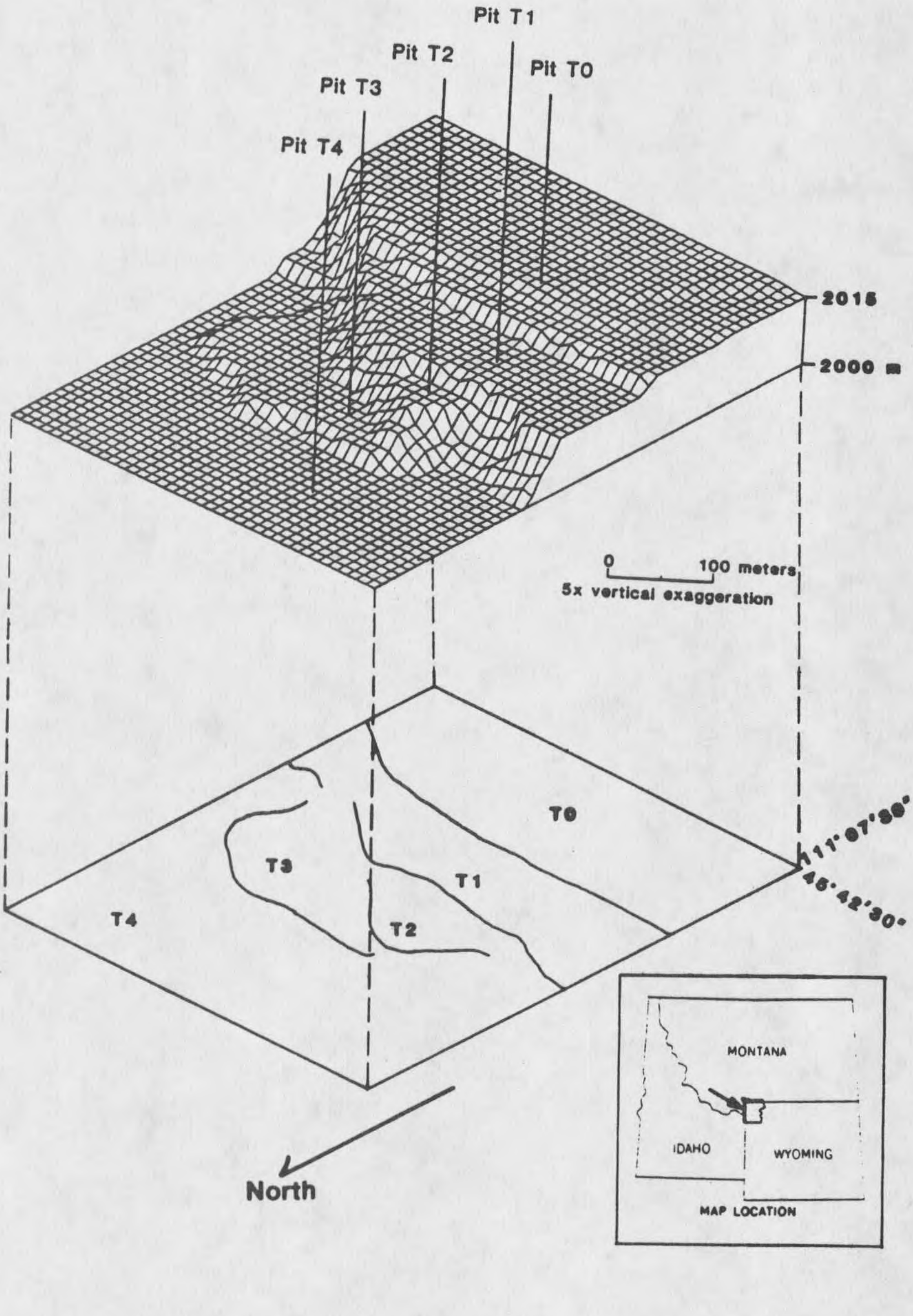


Figure 3 - Computer generated (Golden Graphics, 1984) diagram of study site 1 showing location of sample pits. Sketch map after Nash (1984).

Yellowstone Lake

Sampling pits were dug by the author and William Locke on a sequence of beach bars along a low gradient depositional coastline near Breeze Point for study site 2 (Figs. 2 and 4). This type of coastline was chosen because the material that comprises the bars must have traveled a longer distance than beach material found in steep, erosional coastlines (Komar, 1976). On a low gradient coastline, the features are largely constructional and the material was probably transported from another location while the beach material on a steep coastline may be derived from the adjacent cliff. This difference is important because the beach material is more likely to be reworked if it is transported a greater distance.

Sample pits were dug on the crests of beach berms (Fig. 4) to a depth of one meter and samples collected at 20 cm increments. Soil development was generally poor, but there was a significant amount of organic matter accumulated in the A horizons of some of the pits.

Hydration Rind Measurement

Laboratory procedures used for sample preparation were modified from Michels and Bebrich (1971) (Appendix A). The main difference is that six obsidian grains instead of a single wedge cut from an anthropological artifact are mounted directly to a slide.

All samples were measured blind, in the sense that samples were relabeled by William Locke before preparation and measurement. The identity of the samples was revealed only after measurements were completed. This blinding procedure should decrease the subjectivity of

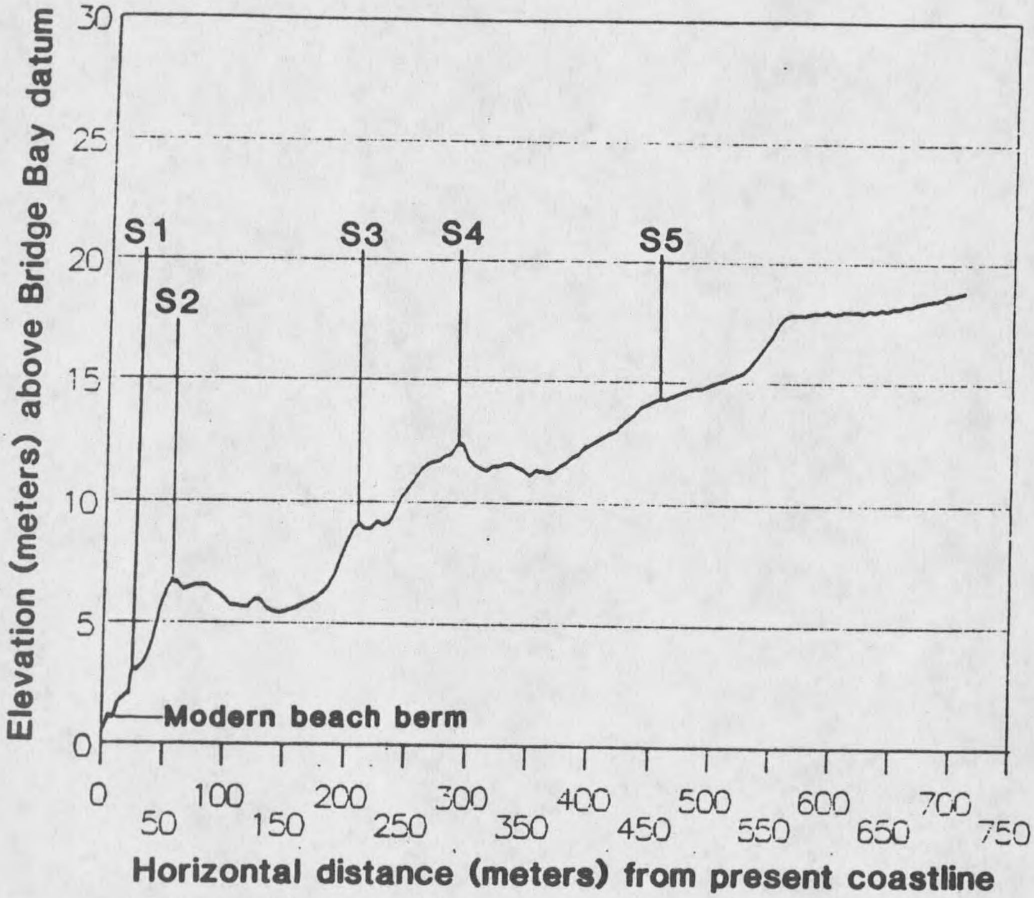


Figure 4 - Surveyed profile of beach berms at Breeze Point showing locations of sample pits. Vertical exaggeration is x20.

the obsidian hydration dating technique as used in this study to insignificant levels.

The optical system used to measure hydration rinds in this study consists of an American Optical polarizing microscope outfitted with a Vickers image-splitting measuring eyepiece. A 10x dry lens was used for measurement coupled with a 15x eyepiece, thereby providing a total magnification of 150x. Admittedly, this magnification is considerably less than what is recommended by Michels and Bebrich (1971), but because statistically valid sample sizes were used, it was thought that a dry lens would be more efficient in terms of time than a higher power oil immersion lens. Furthermore, use of the recommended optical system was limited by what was available in the Department of Earth Sciences at Montana State University.

When viewing individual grains under the microscope each rind measured must first meet a set of criteria which were chosen in order to obtain the most consistent and accurate measurements between grains. These criteria include: 1) smooth edge of grain at least 0.15 mm long 2) edge of grain perpendicular to microscope stage and 3) hydration front parallel to edge of grain and perpendicular to microscope stage. If the measuring location meets the first two criteria but there is no rind present, then a zero thickness rind is recorded. A detailed flow chart for hydration rind measurement procedures is located in appendix B.

The error that is associated with the obsidian hydration dating technique stems from two main sources, which are the resolution of the optical system and measurement error. These sources of error do not vary significantly from one age determination to another (Michels and

Bebrich, 1971). Because the combined operator and instrument error affect all samples in a given study equally, it is recommended by Michels and Bebrich (1971) that each laboratory compute a standard error that applies to all hydration measurements by that laboratory.

The resolution of an optical system is given by:

$$R = 0.610y/NA$$

where R is the limit of resolution (microns), y is the wavelength of light being used (microns) and NA is the numerical aperture of the objective lens (Leach, 1977). The light source used in this study is a tungsten halogen bulb which emits the full spectrum of visible light and corresponds to a wavelength ranging from 0.4 to 0.75 microns. These lower and upper limits are designated y_1 and y_2, respectively.

For the MSU/NPS system the limits of resolution are given by:

$$R_1 = 0.610(0.4)/0.25$$

$$R_1 = 0.976 \text{ microns (violet light)}$$

and

$$R_2 = 0.610(0.75)/0.25$$

$$R_2 = 1.83 \text{ microns (red light)}$$

However, by using an image-splitting measuring eyepiece, the resolution of an optical system can be increased by a factor of ten or more (Dyson, 1960). The final error of the optical system using the image-splitting eyepiece is given by:

$$E = cR$$

where E is the error (microns), c is a constant related to the contrast from the center to the edge of an image and R is the optical resolution of the microscope as given above (Leach, 1977). Values of c generally vary from about 0.096 to 0.068 depending on the general conditions of observation and correspond to an improvement of from 10.4 to 14.7 times the theoretical resolving power of the microscope (Dyson, 1960).

The final theoretical error range of the optical system used in this study is given by the following:

$$E_1 = 0.096(0.976)$$

$$E_1 = 0.094 \text{ microns}$$

and

$$E_2 = 0.096(1.83)$$

$$E_2 = 0.176 \text{ microns}$$

This range compares favorably with the standard error of $\pm$ 0.1 micron accepted by Michels and Bebrich (1971) for their laboratory.

An experiment was done to determine the error associated with measurement of the hydration rinds. Two rinds were measured in the same place thirty times each and the mean and standard deviation computed for each group of measurements (Table 1). The first group has a mean of 5.76 microns and standard deviation of 0.3 microns while the second group has a mean of 19.43 microns and standard deviation of 0.44 microns. It is concluded from this experiment that smaller rinds have a smaller measurement error associated with them. This is consistent with observations in that thicker rinds generally have an indistinct hydration front and so measuring precision would be expected to be somewhat lower.

A standard error of $\pm$ 0.3 microns was accepted because the hydration rinds associated with the fluvial terraces and lakeshores in this study are closer to the first group of error determination measurements than the second (Table 1). The standard error of $\pm$ 0.3 microns also

Table 1. Data array and analysis of measurement error experiment. Two individual hydration rinds were measured thirty times each to determine error associated with measurement technique.

Sample	5-5-1, grain 5	L1-7-3, grain 3
	5.26 microns	18.86 microns
	5.60	19.83
	5.71	20.51
	5.66	20.06
	5.49	19.09
	5.89	19.54
	5.89	20.00
	5.37	20.00
	5.71	19.03
	5.71	19.77
	5.66	19.66
	6.00	19.43
	5.77	18.91
	5.60	19.31
	5.83	19.60
	5.43	19.83
	6.06	19.60
	5.66	19.43
	6.06	19.37
	5.83	19.14
	6.17	19.43
	6.11	19.20
	6.29	19.37
	5.09	19.03
	6.17	18.63
	6.29	18.46
	5.83	19.89
	5.60	19.43
	5.77	19.43
	5.26	19.09
Mean	5.76	19.43
Std Dev	0.30	0.44

includes the error associated with the optical system.

Grain Size Selection

All of the samples from both the fluvial terraces and the lake-shores contain sediment size classes from fine sand through coarse gravel. Therefore, an experiment was performed to determine the optimal grain size to work with in this study, both in terms of the number of measurements taken per given number of grains and ease of sample preparation. The sample chosen for this experiment (KDA-88-TO, 80-100 cm) was obtained from the OSP. The 80 to 100 cm level was chosen because it is the least likely to have been affected by bioturbation and solar insolation. Sample grain sizes range from coarse silt to pebble size particles (0.0625 to 16 mm), but not all of these sizes were prepared for measurement because of laboratory limitations. The sizes that were prepared are as follows: 0.5 to 1 mm, 1 to 2 mm, 2 to 4 mm and 4 to 8 mm. Slides were prepared with at least 30 grains for each size class and examined under the petrographic microscope to determine hydration rind thicknesses (Table 2).

The sample means of the hydration rind thicknesses are statistically indistinguishable for the sample sizes examined (Table 2). In addition, each size class examined has a comparable range of rind thicknesses, therefore it was assumed that the size class that offered the greatest number of measurements per given number of grains would be the most efficient to work with for the remainder of the study.

From Table 2 it is concluded that the optimum grain size to prepare for hydration rind measurements is the 2 to 4 mm class because of the

Table 2. Data array and analysis of hydration rind thicknesses for different grain size classes from Terrace T0 (80-100 cm).

	Size Classes			
	0.5 - 1 mm	1 - 2 mm	2 - 4 mm	4 - 8 mm
	4.06	4.44	2.83	5.28
	5.28	5.33	5.50	6.72
	7.39	7.06	6.17	7.72
	10.78	7.83	6.39	9.28
	14.61	8.78	7.72	9.44
	16.89	9.00	8.44	9.56
	19.06	9.72	9.94	10.67
	19.17	10.56	10.11	11.33
	21.89	12.00	11.89	11.61
	23.56	12.39	12.22	12.72
		12.50	12.44	13.72
		13.72	13.50	13.78
		14.61	14.00	14.28
		17.06	14.11	14.44
		17.22	14.56	14.50
		18.94	15.00	16.06
		20.78	19.67	16.11
		22.67	20.28	16.17
		24.33	20.39	17.44
		24.61	20.61	17.72
			21.22	18.11
			21.50	20.17
			22.39	22.83
			23.17	23.17
			25.33	
			25.89	
			27.94	
N	n=10	n=20	n=27	n=24
Mean	14.27	13.68	15.30	13.87
Std Dev	7.01	6.19	6.98	4.72
Std Error	2.22	1.38	1.34	0.96

number of measurements taken per number of grains. Although the 4 to 8 mm class showed the smallest standard error of the mean (SEM), the difficulty of preparation of the larger grains dictated the choice of the 2 to 4 mm class. This smaller class size is thought to represent a balance between high energy depositional environments and ease of sample preparation. Grains larger than this are more difficult to work with and take longer to prepare, but presumably are representative of higher energy depositional environments (Church, 1972). Smaller grains are somewhat easier to prepare, but did not have as many measurable rinds for a given number of grains. The liklihood of reworking for smaller grains is not as great as with larger grains due to the cushioning effect of a water film surrounding each grain (Ziegler, 1911; Thiel, 1940). Smaller grains are also more likely to travel in suspension which is not an effective reworking mechanism (Dobkins and Folk, 1970).

In hindsight, the choice of sample KDA-88-TO, 80-100 cm was unfortunate in that it is one of the few bimodal samples encountered in this study (see RESULTS). Thus the tentative conclusion of comparable rind development on different sized grains has not been adequately tested.

Zero Thickness Hydration Rinds

To determine if obsidian gravel was effectively reworked in the fluvial system an experiment was performed to test the hypothesis that the ancestral Madison River sufficiently reworked obsidian grains so that each terrace possesses a unique, characteristic hydration rind thickness. It was assumed for this experiment that the modern stream serves as an appropriate analog for the ancestral Madison River which cut the terraces. The experiment consisted of examining obsidian grains from Terraces T1 to T4 and the modern stream and specifically noting the presence of zero thickness rinds found on each grain. Many grains had more than one location where a zero thickness rind could be observed, as outlined by the measurement site selection criteria, but not more than one zero thickness rind was recorded for each grain. By classifying each grain as possessing zero rinds or not possessing zero rinds, the statistical tests are more significant (Richard Rossi, 1989, personal comm.)

The data were analyzed using a series of 2 x 2 chi square tests (with a 5 % rejection level) to test whether there was a difference in the number of zeros found on each of the upper terraces compared to the modern stream (Tables 3 and 4). The first hypothesis is that the number of zero thickness rinds on all surfaces is equal. This hypothesis was rejected with a P-value = 0.0135 (Table 3): there is only a 1 % chance of this observed difference in zero values having arisen from sampling of similar populations. The rejection of this hypothesis indicates that there is a difference between samples.

The second hypothesis is that the modern stream is similar to the four other surfaces with respect to the number of zero thickness rinds. This hypothesis was rejected with a P-value = 0.0032 (Table 4) which indicates that there is a difference between the modern stream sample and Terraces T1 through T4.

The results of these two statistical tests imply that reworking does take place in the modern stream. However, reworking is not complete

Table 3. Chi square test to determine if all terraces have same number of zero thickness rinds.

	Rind Thickness	
	zero	non-zero
Terrace T1	15	20
T2	11	24
T3	14	22
T4	20	17
Modern stream	25	11

H_o: T1 = T2 = T3 = T4 = Modern stream
P-value = 0.0135
Therefore reject H_o, there is a difference between samples.
Note: 35 to 37 grains were measured from each surface with each grain having only one measurement.

Table 4. Chi square test to determine if the modern stream has the same number of zero thickness rinds as the terraces.

	Rind Thickness	
	zero	non-zero
Terraces T1 to T4	60	83
Modern stream	25	11

H_o: T1 + T2 + T3 + T4 = Modern stream
P-value = 0.0032
Therefore reject H_o, there is a difference between the terraces and the modern stream.

as indicated by the presence of hydration rinds on gravel sampled from the modern stream. This is consistent with the observation of Pierce and others (1976) that the grains they examined for evidence of glacial reworking also possessed preexisting rinds that dated from earlier events. Therefore, each reworking event does not completely erase the record of previous reworking events, potentially enabling several events to be dated from a single grain.

Statistical Methods and Data Presentation

Hydration rinds on obsidian artifacts are clearly identifiable and present little ambiguity in correlating a rind with the creation of the artifact (Friedman and Smith, 1960). However, when examining rinds from naturally worked obsidian, the rinds that date from the event of interest must be selected from a number of rind thicknesses. Pierce and others (1976), in their study of glacially worked obsidian, reasoned that both radial pressure cracks and shear cracks on obsidian gravel stemmed from glacial processes and so could concentrate their measurement efforts on rinds that formed along these cracks.

The lack of clearly identifiable event-specific hydration rinds to measure in the case of fluvial and lacustrine worked sediments dictates a statistical approach to the problem. As mentioned above, on a given grain there may be several rind thicknesses and therefore it is logical to assume that the thinnest rinds date from the most recent event and that progressively thicker rinds date from older events. However, because the hydration phenomena causes mechanical strain, rinds tend to spall parallel to the grain edge (Michels and Bebrich, 1971) and thus may cause anomolously thin rinds to be present. Consequently, sample sizes must be sufficiently large to be able to differentiate between rinds that date from the event of interest and those that are created by other processes (i.e. spalling and frost fracture).

A further reason statistically valid sample sizes must be used for naturally worked specimens is that it is likely that hydration rate varies within individual rhyolite flows and is generally accepted that hydration rate varies between flows (Friedman and Long, 1976; Michels, 1985; Michels, 1988). Friedman and Long (1976) postulated relationships between silica content and hydration rate and between refractive index and hydration rate. They also found that increased CaO (calcium oxide) and MgO (magnesium oxide) reduce the hydration rate. Analyses of single-source (Aster Creek and Obsidian Cliff flows) obsidian shows variation among the chemical constituents of individual samples (Appendix C) (Friedman and Long, 1976; Michels, 1988). In particular, the amounts of CaO and MgO in the Aster Creek flow vary by as much as factors of 1.4 and 2.0, respectively. The analyses indicate that these flows probably do not have a unique hydration rate but that rates are variable around some representative hydration rate. Therefore, single-source obsidian grains may possess different rind thicknesses that date from the same event.

Furthermore, because each obsidian flow has a representative hydration rate (Friedman and Long, 1976), deposits with more than one source area may also possess rinds of different thicknesses that date from the same event. If a given set of sequential deposits such as fluvial terraces contain the same mixture of source material, statistically valid sample sizes from each terrace should alleviate the problem of differing hydration rates, assuming that each terrace contains the same mixture of source material.

Nonparametric density estimation (Tarter and Kronmal, 1976) was used to display the rind measurement data from the sample pits on the fluvial terraces and lakeshores. Density estimation assigns each data point a standard distribution and the sum total of all data points defines the shape of the overall density estimation curve. However, the shape of the curve can be slightly altered by choosing different smoothing parameters. The advantages of density estimation over parametric frequency histograms is that bias is not introduced by choosing a class interval and the technique does not assume that the sample was taken from a normally distributed population. Furthermore, density estimation was used because the data distribution for each of the samples appears to be a mixture of several normally distributed populations (Richard Rossi, 1989, personal comm.). This is a situation which does not easily lend itself to parametric statistical analysis. However, curve fitting procedures are currently in progress that will allow parametric estimates of the data distributions (Richard Rossi, 1990, personal comm.).

RESULTS

Obsidian Sand Plain

Rind measurement data, exclusive of zero rinds, for individual sampling levels for Terrace T0 (Appendix D) are shown by density estimation diagrams in Figures 5-9. Hydration rinds from each sampling depth were measured in order to determine if hydration rind thickness varied with depth and to define an appropriate depth from which to measure rinds in the remainder of the sampling pits. The 20 to 40 cm and 40 to 60 cm levels for Terrace T0 (Figs. 6 and 7) show the most regular peak in the 0 to 10 micron range and therefore this range of levels was used in all subsequent sample measurements.

It should be noted that all of the different pit levels except the 80 to 100 cm level show a similar pattern of data distribution (Figs. 5-9). The data for the upper 80 cm of the pit suggest that the hydration rinds record at least three events with the youngest centered around 6.5 microns. The dominant peak at about 6.5 microns for the 80 to 100 cm depth is much subdued and the distribution instead displays strong peaks at about 14 and 22 microns (Fig. 9). The aggregate data for Terrace T0 clearly shows all three peaks (Fig. 10).

Rind measurement data, exclusive of zero rinds, from the 20 to 60 cm depth for Terraces T1 to T4 and surface samples for the modern stream (Appendix E) are shown by density estimation diagrams in Figures 11-15. These density diagrams show a strong peak centered around 6.5 microns.

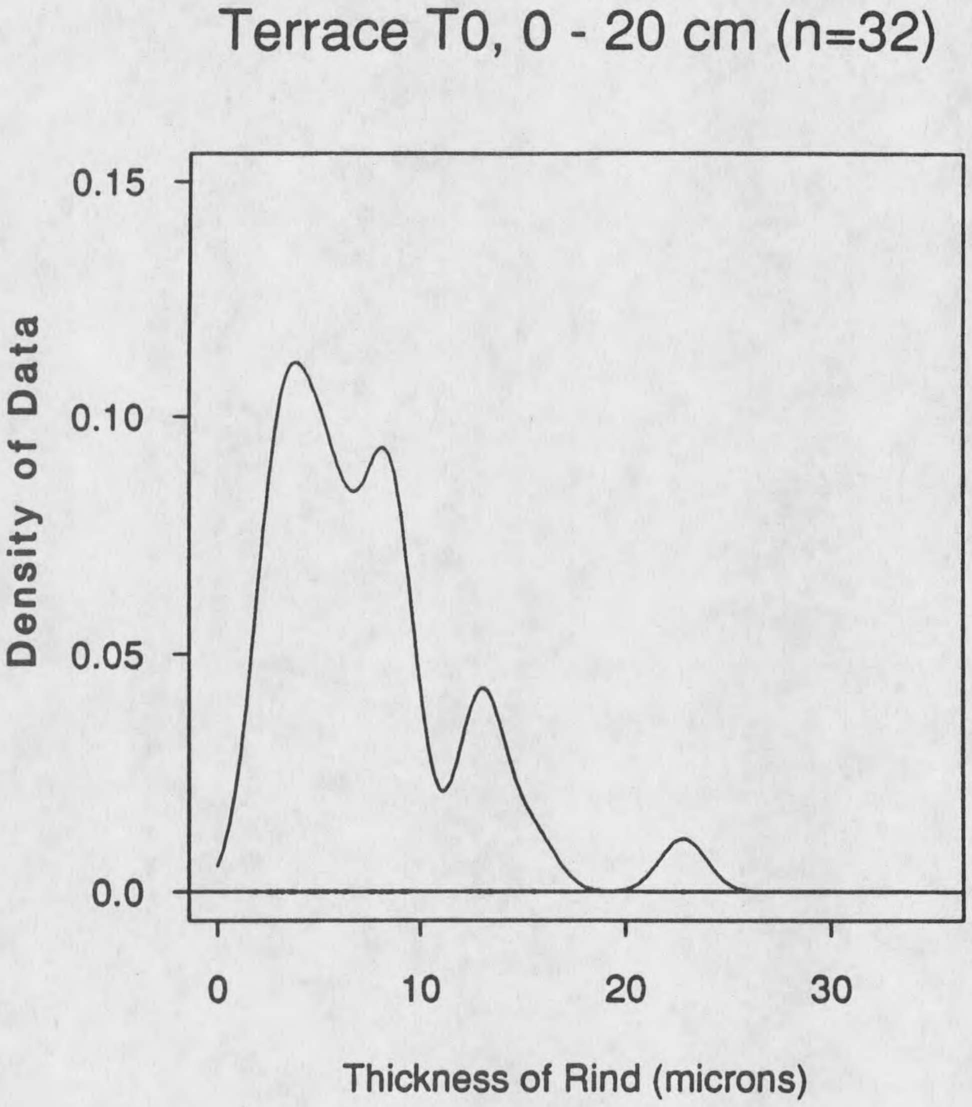


Figure 5 - Rind measurement data for Terrace T0 at 0 - 20 cm depth.

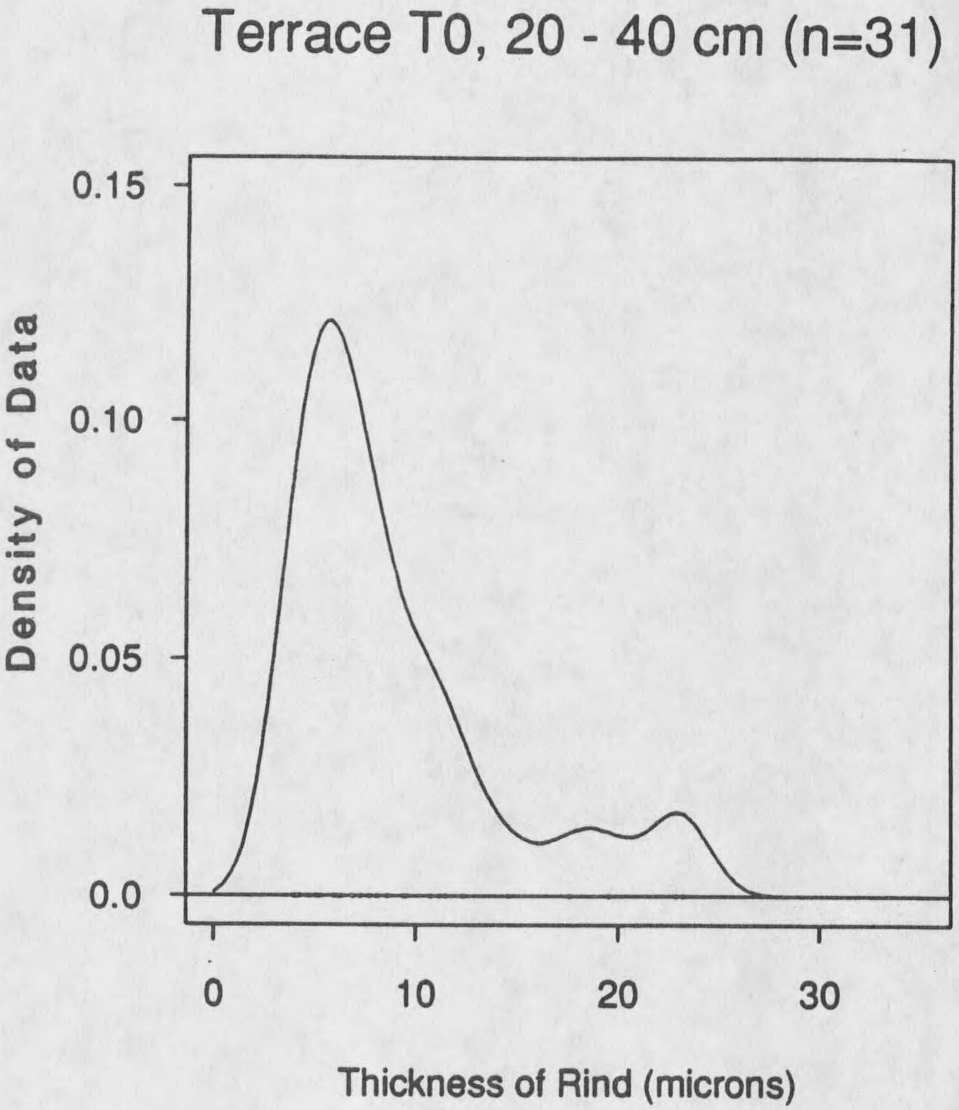


Figure 6 - Rind measurement data for Terrace T0 at 20 - 40 cm depth.

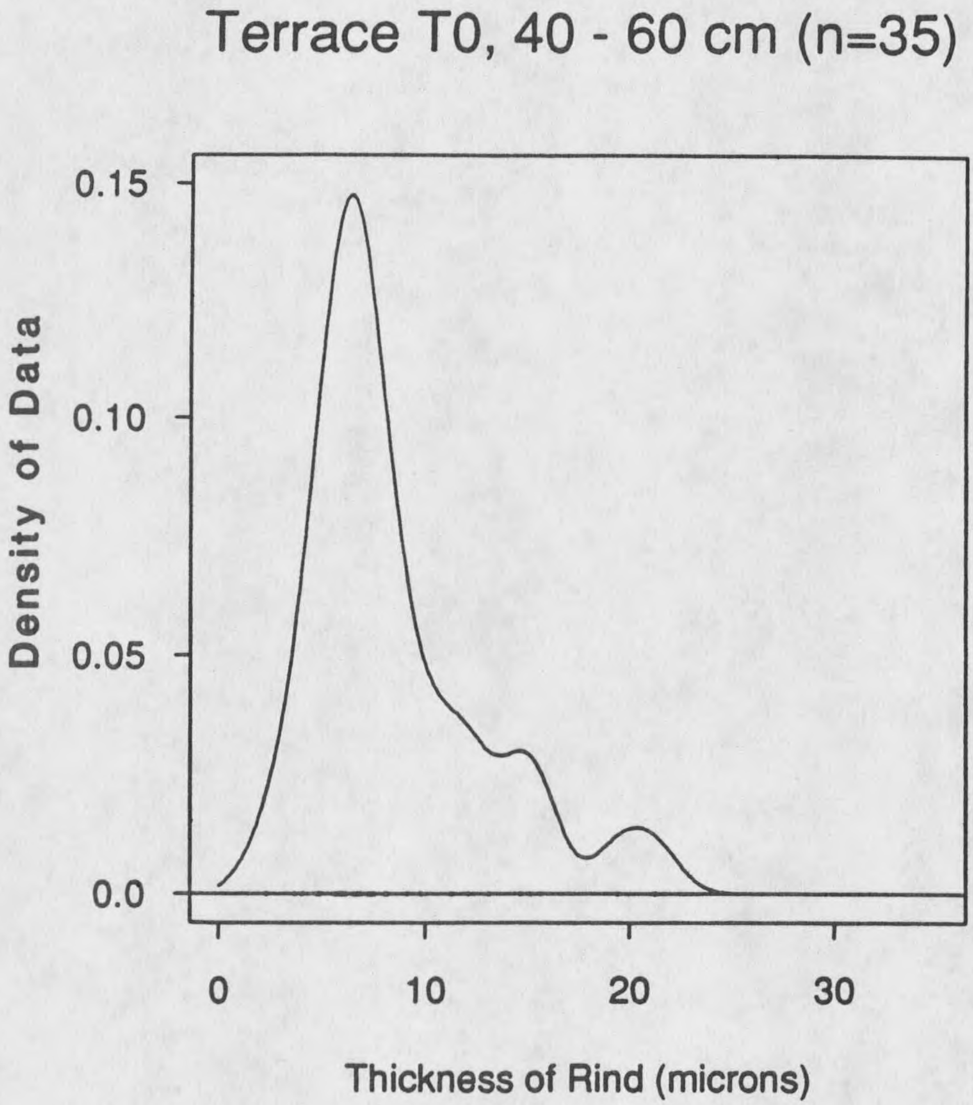


Figure 7 - Rind measurement data for Terrace T0 at 40 - 60 cm depth.

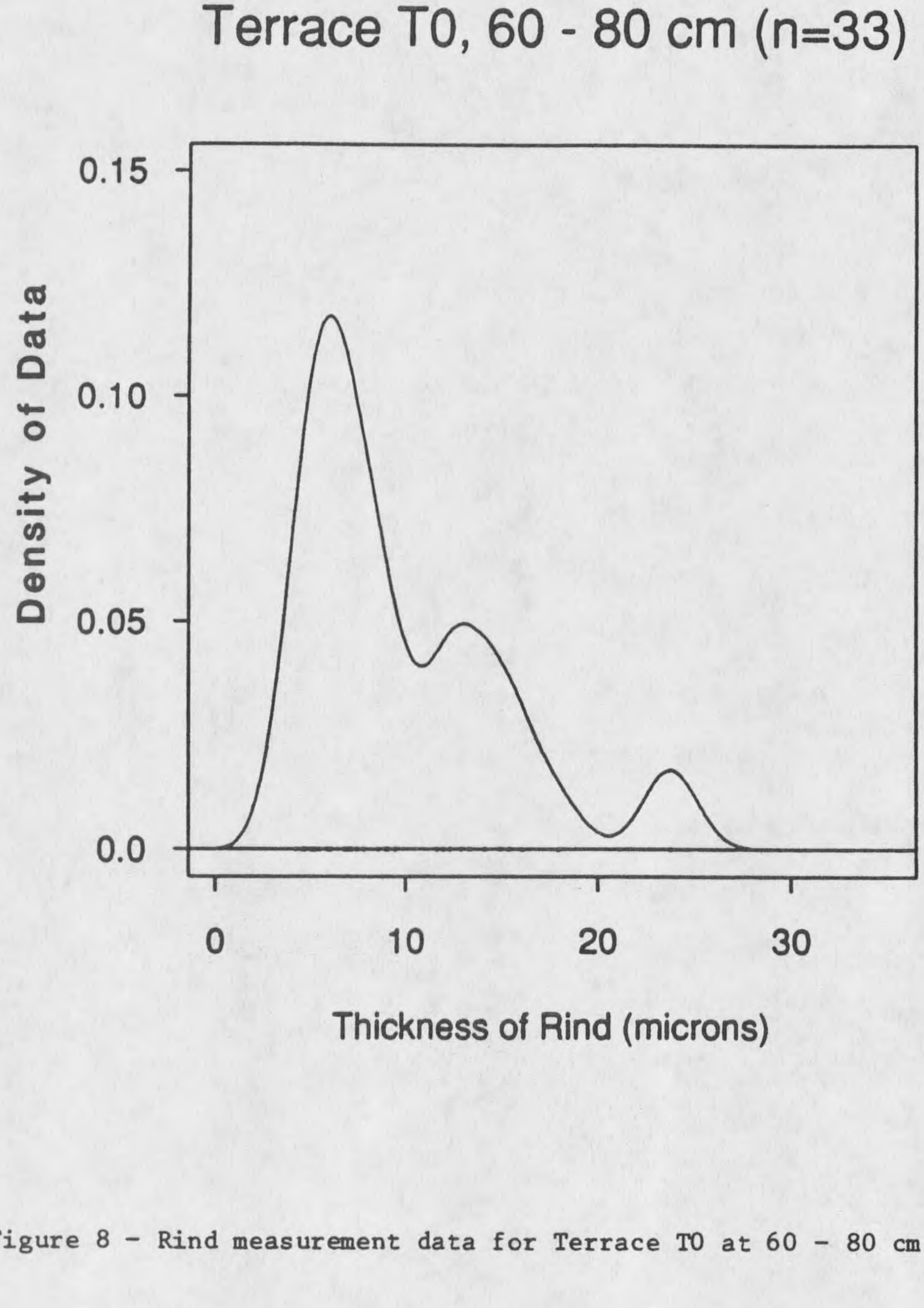


Figure 8 - Rind measurement data for Terrace T0 at 60 - 80 cm depth.

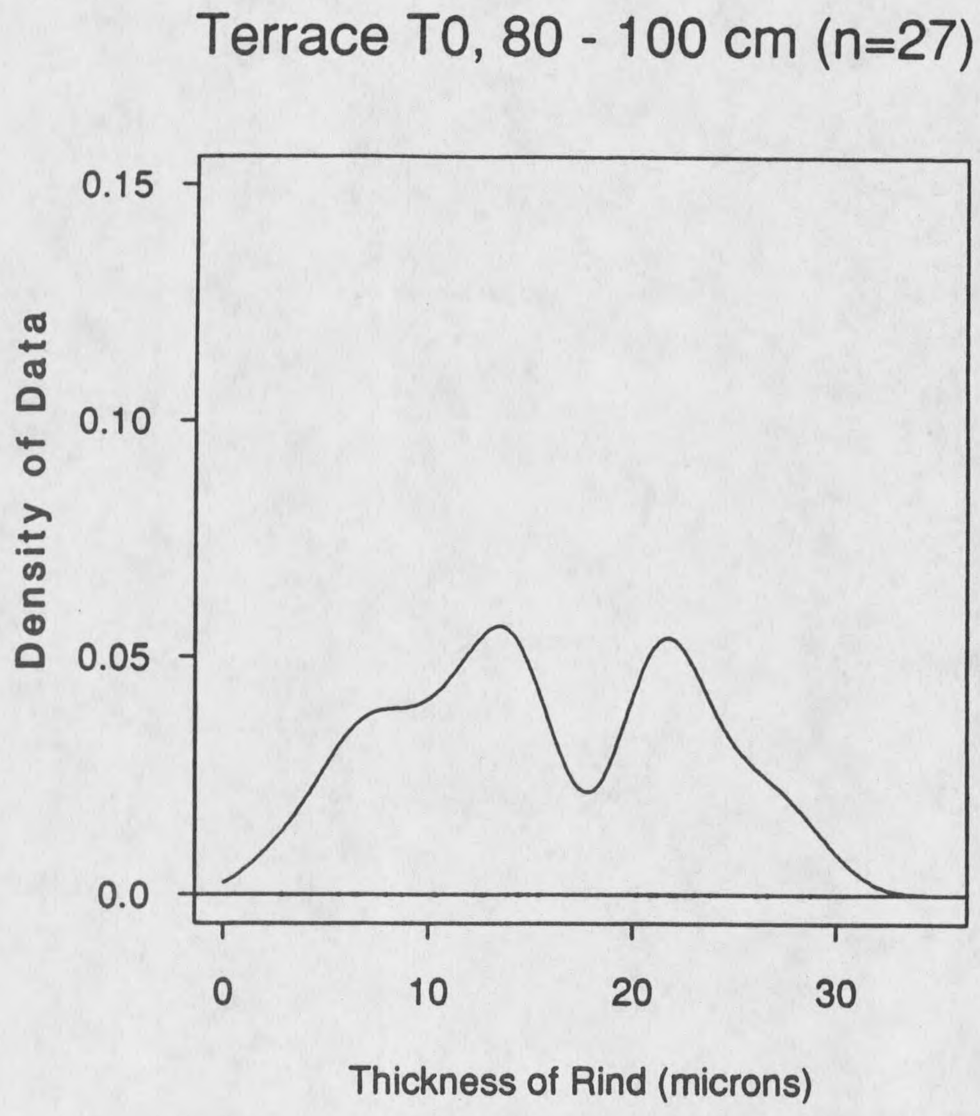


Figure 9 - Rind measurement data for Terrace T0 at 80 - 100 cm depth.

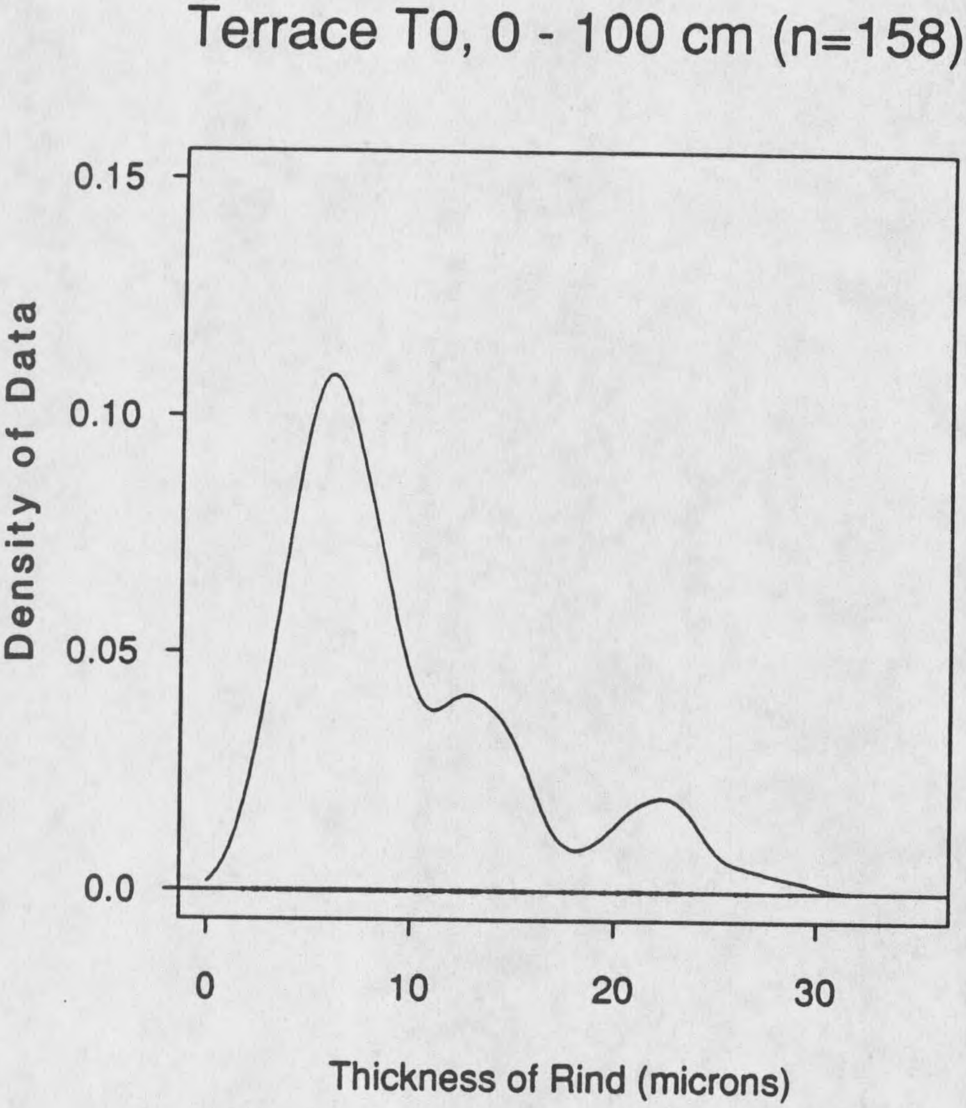


Figure 10 - Aggregate rind measurement data for Terrace T0.

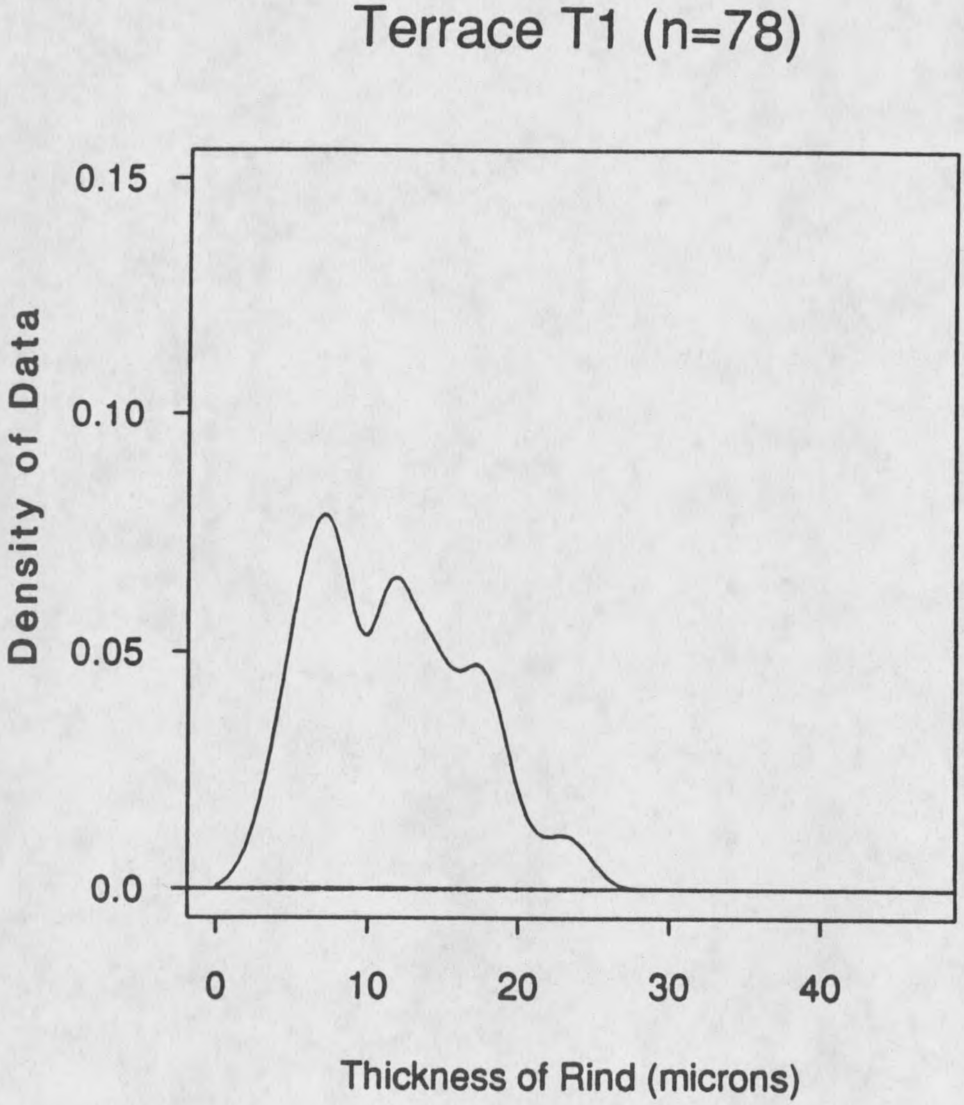


Figure 11 - Rind measurement data for Terrace T1.

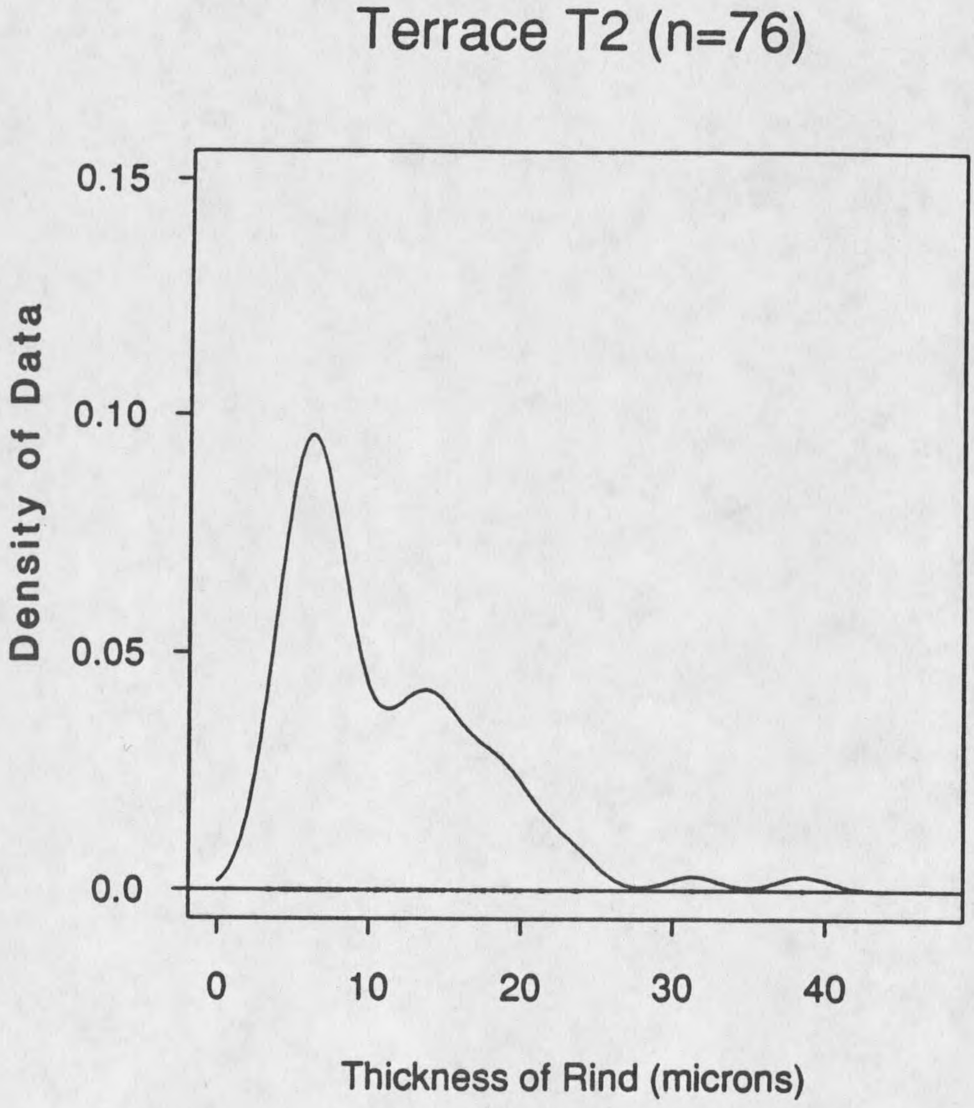


Figure 12 - Rind measurement data for Terrace T2.

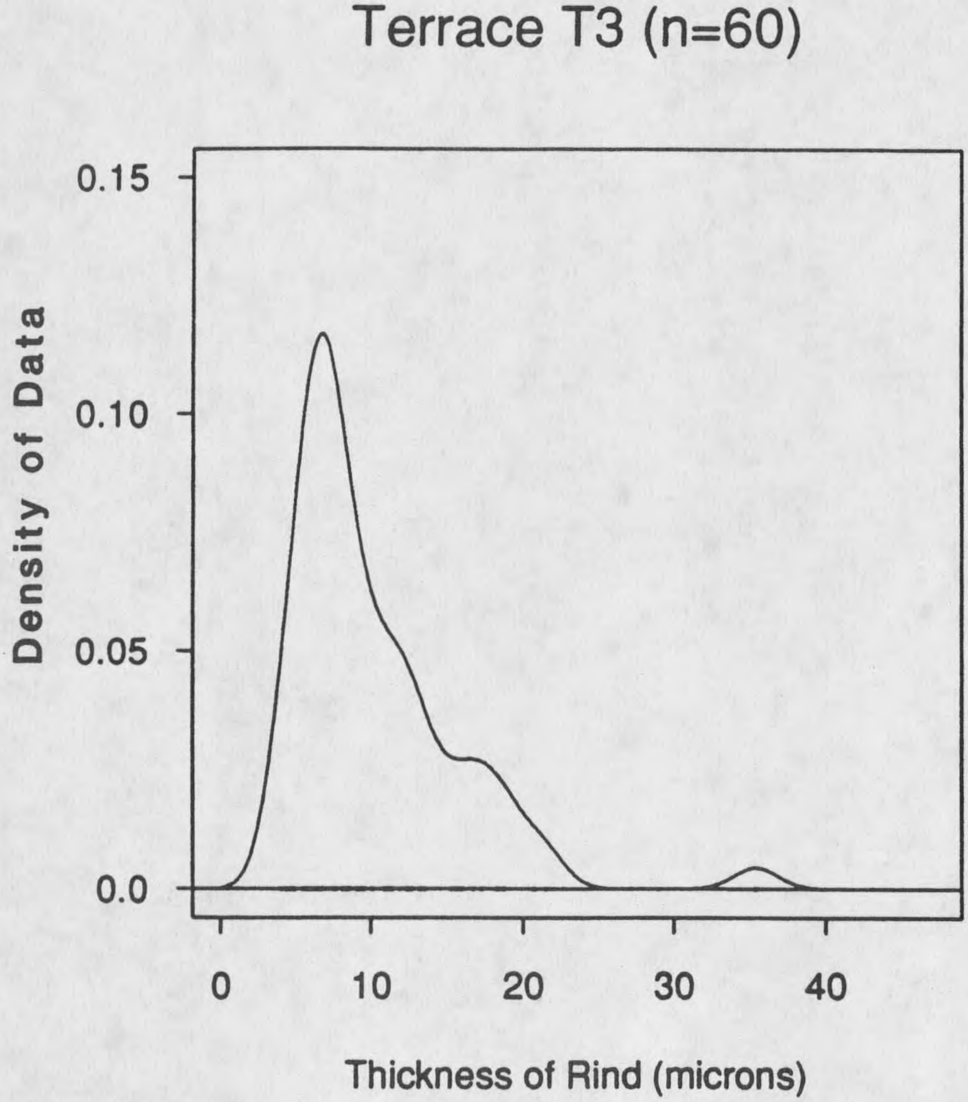


Figure 13 - Rind measurement data for Terrace T3.

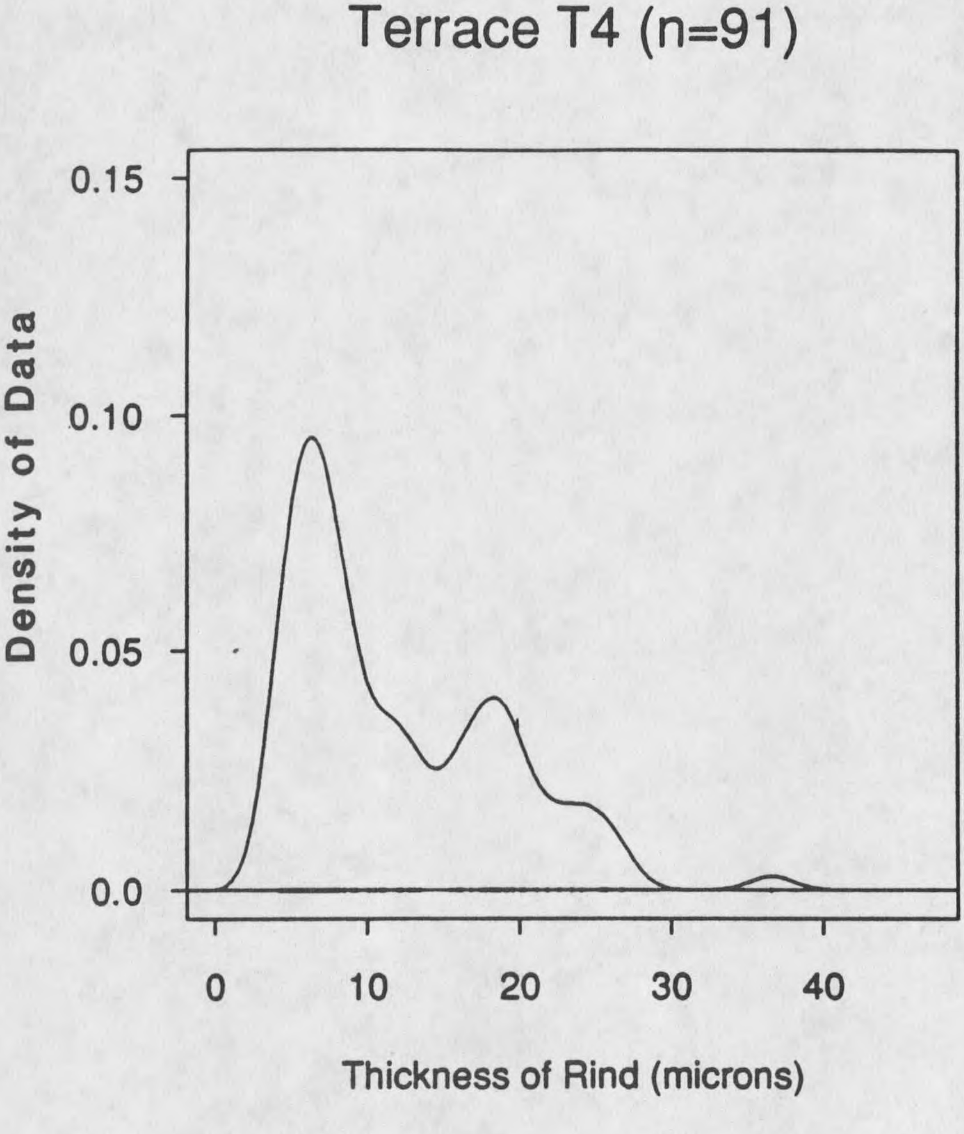


Figure 14 - Rind measurement data for Terrace T4.

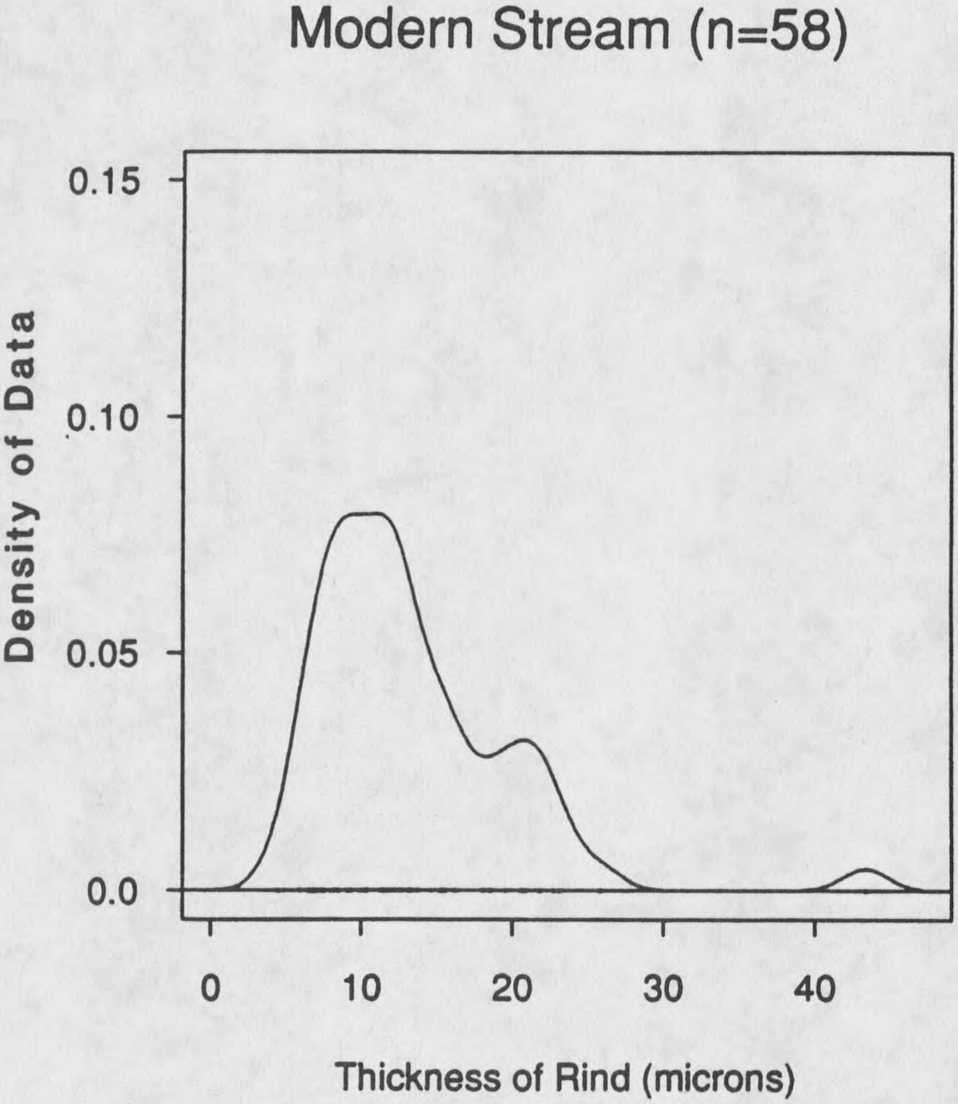


Figure 15 - Rind measurement data for modern stream.

However, the distribution of rinds thicker than 6.5 microns for Terraces T1 to T4 and the modern stream do not appear to consistently cluster.

Yellowstone Lake

Rind measurement data, exclusive of zero rinds, for the five raised shorelines (Appendix F) are shown by density estimation diagrams in Figures 16-20. Each of the diagrams has a prominent peak between 5 and 6.2 microns. Rinds thicker than this do not appear to cluster around common points for all shorelines but are distributed between about 10 and 30 microns. Both Shorelines S3 and S4 (Figs. 18 and 19) also show a fairly prominent peak at about 16 microns. The uppermost shoreline, S5, displays a peak at about 21 microns that is as prominent as the peak at 5.2 microns (Fig. 20).

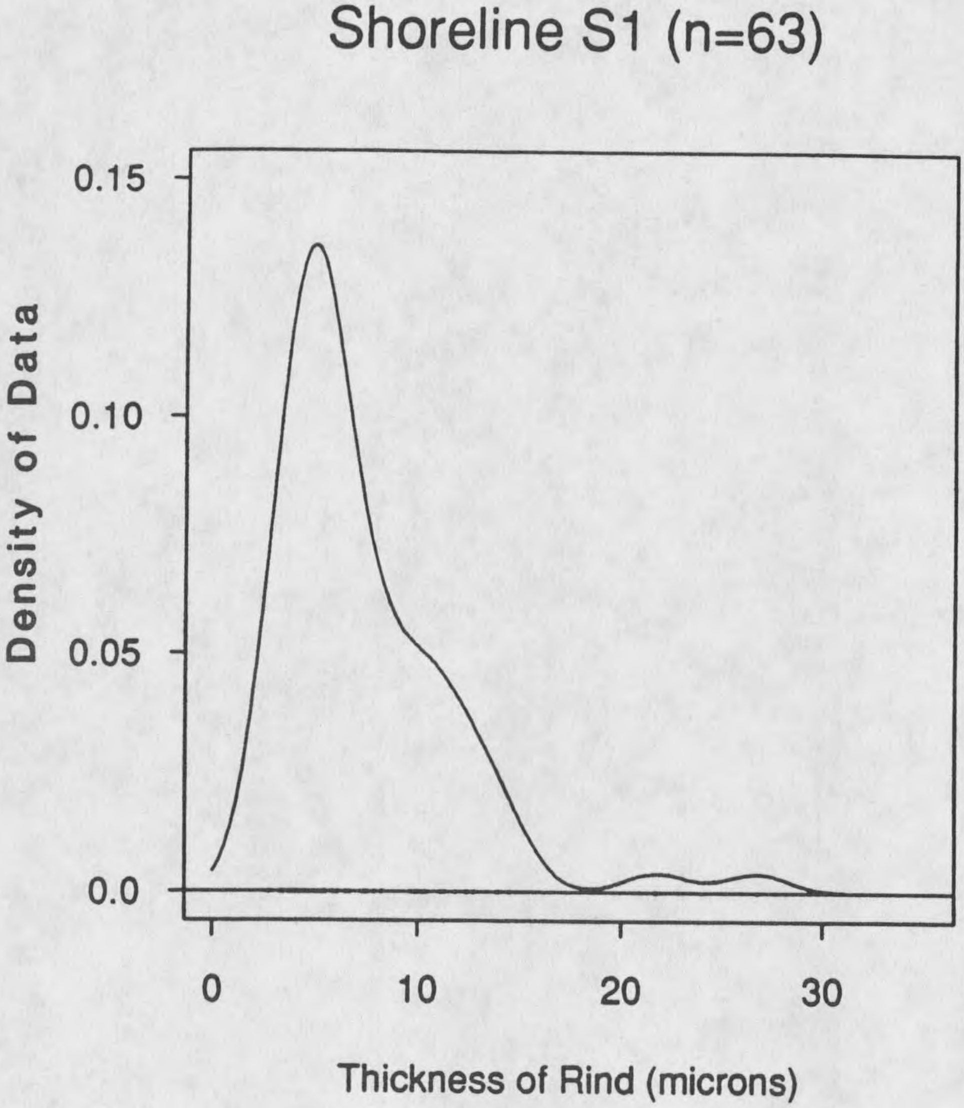


Figure 16 - Rind measurement data for Shoreline S1.

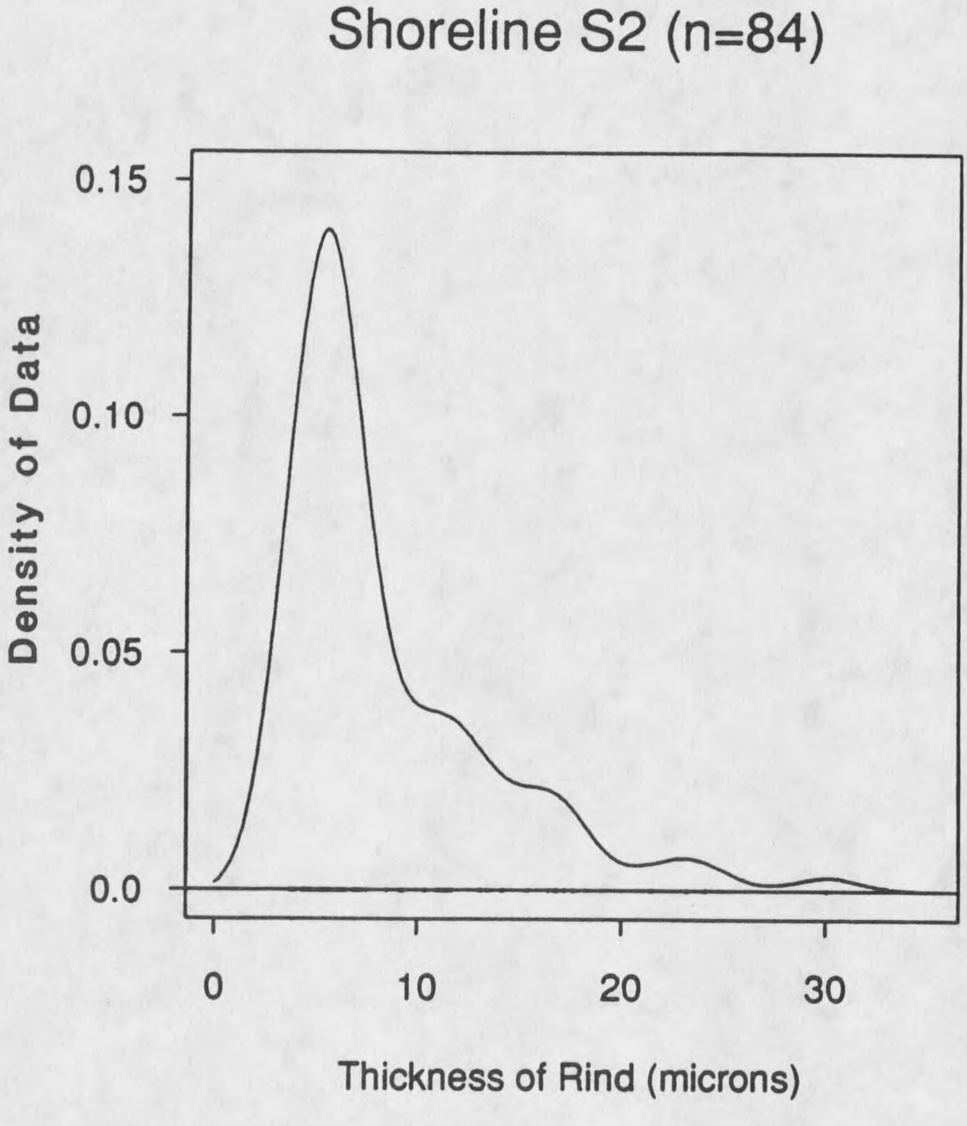


Figure 17 - Rind measurement data for Shoreline S2.

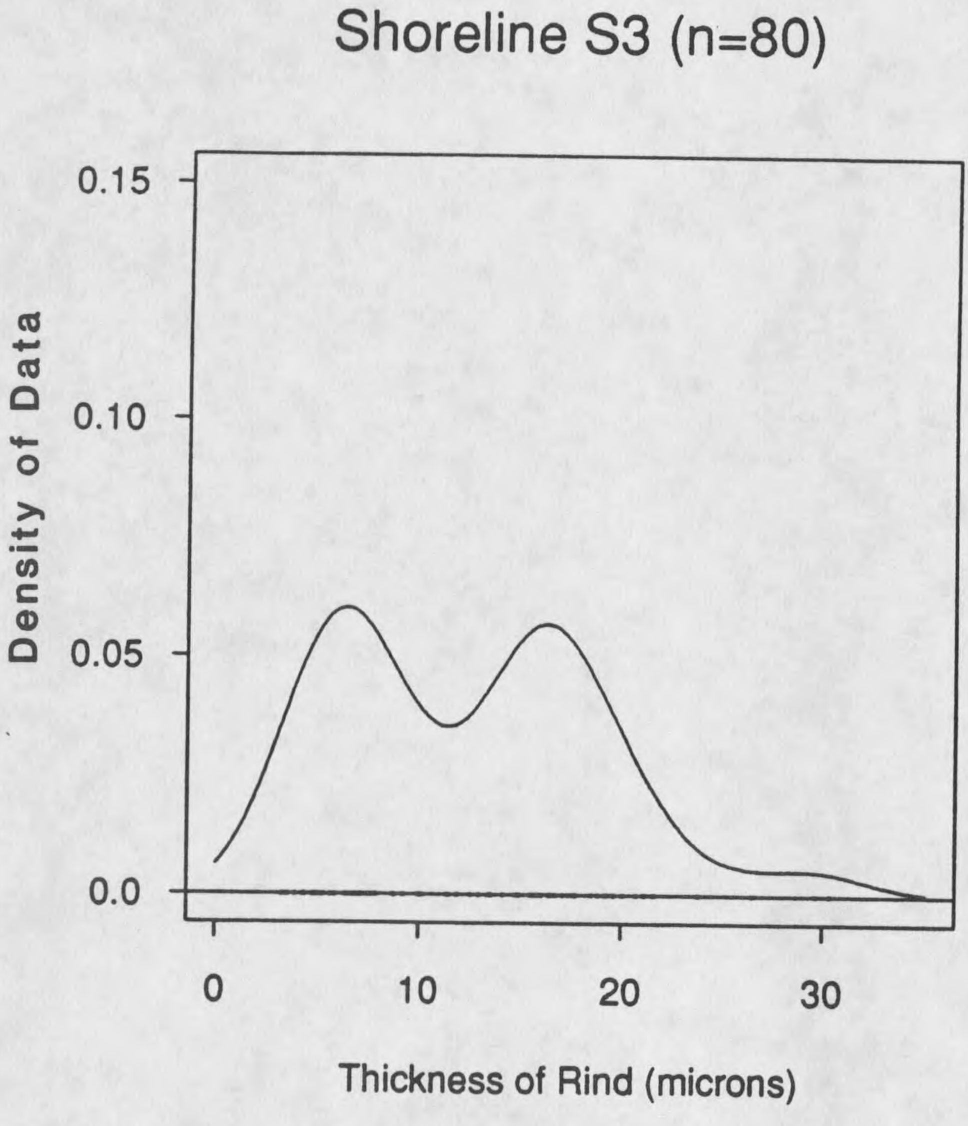


Figure 18 - Rind measurement data for Shoreline S3.

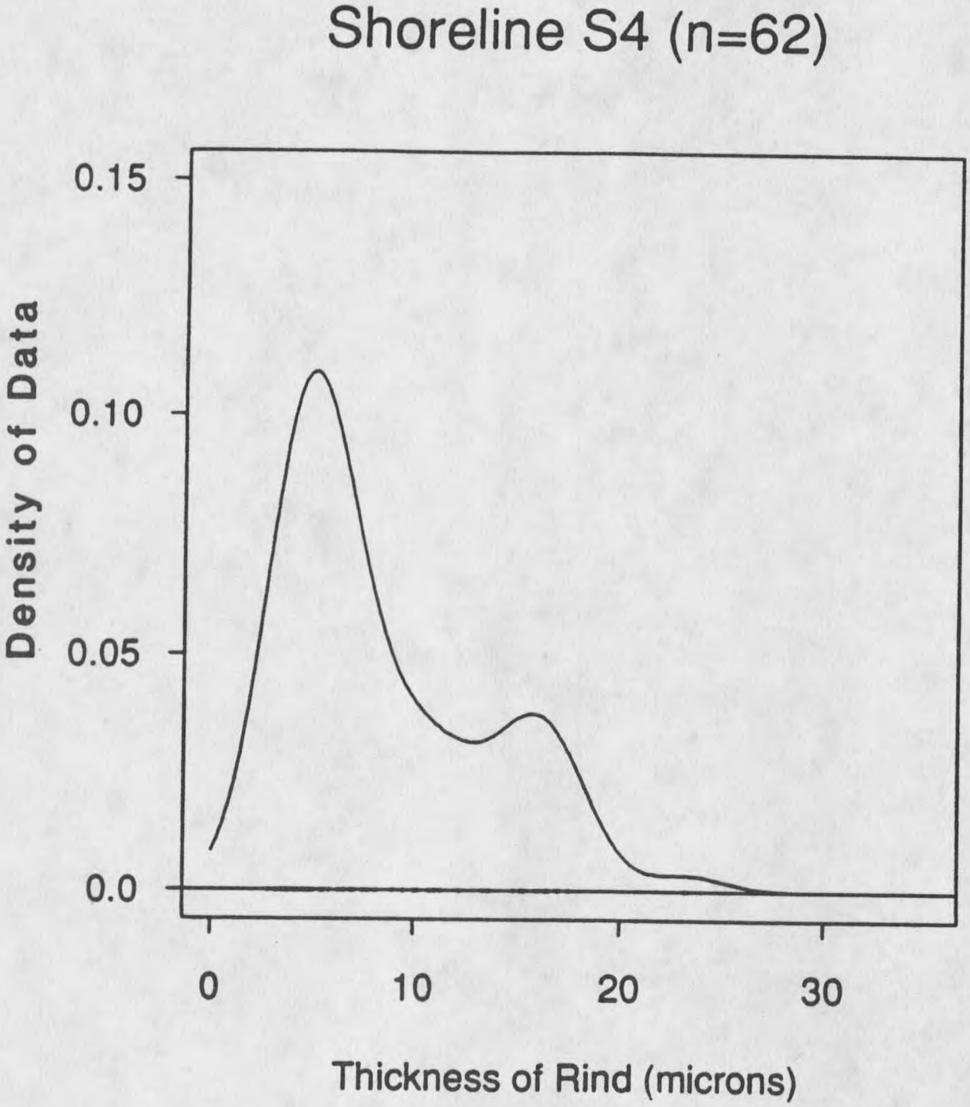


Figure 19 - Rind measurement data for Shoreline S4.

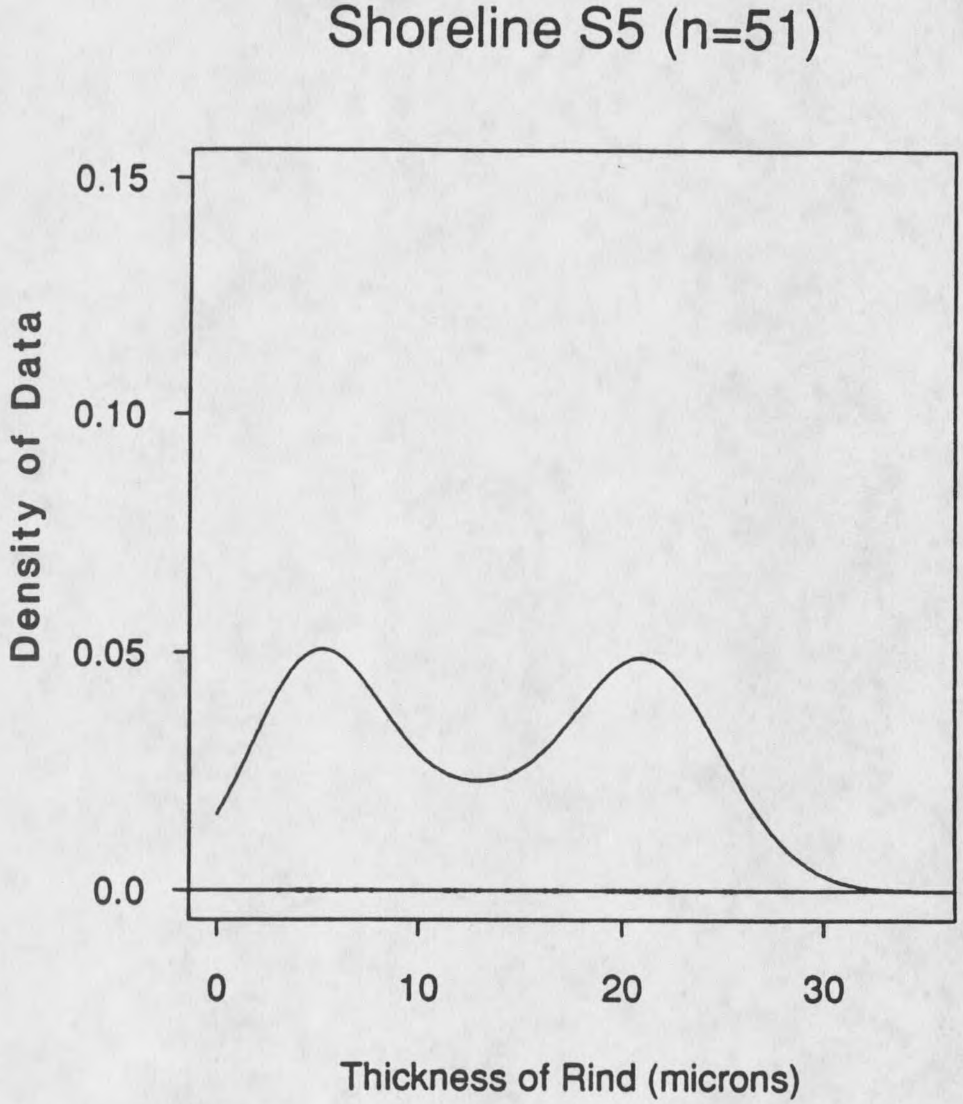


Figure 20 - Rind measurement data for shoreline S5.

DISCUSSION

The original purpose of this study was to devise a method, using obsidian hydration dating techniques, to date fluvial terraces and lake shorelines. During the course of this study, much information has been revealed about the obsidian hydration dating technique in general. This information has important ramifications for geologists and archaeologists employing the technique in their studies. In the following section, the technique will first be discussed as applied to fluvial terraces and lakeshores and then a general discussion of the obsidian hydration dating technique will follow.

Obsidian Sand Plain

Solar Insolation

Friedman and Long (1976) found that obsidian exposed to the sun will hydrate much more rapidly than buried obsidian. In their experiments near West Yellowstone, MT, they determined that obsidian from the Obsidian Cliff flow reached a temperature of 19.7 °C in direct sunlight which corresponds to a hydration rate of 16.0 microns^2/1000 yrs. Soil temperature at a depth of 0.9 m averaged 6.9 °C, which corresponds to a hydration rate of 3.4 microns^2/1000 yrs. The disparity between these two rates is very significant with regard to sampling depth. It would appear from the study of Friedman and Long (1976) that to escape the effects of solar heating, samples should only be collected at some depth below the surface.

The data from Terrace T0 (Figs. 5-9) show that there is no apparent increase in rind thickness in the upper few decimeters which implies that effects of solar heating are negligible at this site. It was expected that there would be an increase in rind thickness toward the surface, but this is not portrayed by the data. One possible explanation is that bioturbation could have the effect of mixing the surface layer thereby nullifying the solar heating signal. Bioturbation may take the form of root throw or ground squirrel burrowing. It is also possible that, because this site is covered by snow for a large part of each year, solar heating does not have an effect on the hydration rate.

Reworking Processes

Reworking in the fluvial environment is probably dominated by fracturing as opposed to abrasion. Twenhofel (1945), in his study of the rounding of sand grains, stated that when stream sediment contains gravel and sand, the sand particles do not become rounded. Instead, the sand grains are repeatedly fractured by impact and grinding. It is not unreasonable to suggest that gravel size particles also remain angular through the same processes of reworking. Dobkins and Folk (1970) reported that stream gravels retain much of their angularity during transport. This implies that if reworking does occur in the fluvial environment, then the dominant process is probably fracturing as opposed to abrasion, which would tend to increase the roundness of individual clasts (Dobkins and Folk, 1970).

When examining hand specimens from the OSP, some of the obsidian pebbles have slightly frosted corners, which probably resulted from abrasion, but generally the obsidian grains are very angular and irregular in shape. When examining these pebbles with a microscope, some of the corners appear abraded, but overall the edges of the pebbles are generally smooth. No rinds were measured on corners that appeared abraded because the rinds in these areas did not have consistant thicknesses along the abraded edge. Instead, all rinds were measured on smooth edges. A smooth edge would be expected to be created through the conchoidal fracture of obsidian. The above observations indicate that obsidian pebbles wear in the fluvial environment mainly by fracturing and hydration rinds were measured only on fracture surfaces.

The modern stream contains significantly more zero thickness rinds than the higher terraces (Tables 3 and 4) which implies that fluvial reworking is an effective agent in resetting rinds on obsidian grains. The modern stream possesses about 20 to 60 % more zero thickness rinds than the other surfaces (Table 3) which indicates that reworking processes are effective on about 20 to 60 % of the grains. The processes that produce zero thickness rinds on Terraces T1 to T4 (Table 3) most probably reflect in situ fracturing, sampling technique and sample preparation.

Grains are only partially reworked in the fluvial environment which implies that each grain can potentially record multiple events. This is evidenced by multiple rind thicknesses for each pit level (Figs. 5-9) and multiple rind thicknesses on individual grains as observed in this study and reported by Pierce and others (1976). Additional evidence includes the fact that grains in the modern stream possess rinds of various thickness in addition to zero thickness rinds (Appendix E). If complete reworking occurs in the fluvial environment, there should not be any measurable rinds present on modern stream gravel and by inference, gravel from an individual terrace should only possess rinds that date from a single event. Both of these situations are clearly not the case.

Depth of Reworking

There was no visible stratigraphy in the pit on Terrace T0, however the data from the 80 to 100 cm depth (Fig. 9) lack the dominant peak at about 6.5 microns and instead show two prominent peaks at about 13 and 22 microns, respectively. These data are not compatible with a Pinedale age for this material and can be interpreted in at least two ways. The first hypothesis is that the 80 cm depth represents maximum depth of reworking by a Pinedale age stream. This implies that the bulk of the deposit dates from an earlier depositional event and Pinedale reworking only affected the upper 80 cm of the OSP at this location.

The second hypothesis is that the amount of material above the 80 cm depth represents Pinedale aggradation. This hypothesis is genetically similar to the first hypothesis but differs in at least one important aspect. Both hypotheses call for some Pinedale-age mechanism to reset grains in the upper 80 cm of the deposit, however one calls for fluvial reworking where the second calls for aggradation, which is compatible with Pierce's (1979) outburst flood hypothesis.

The author proposes that the OSP does not date from one particular episode of deposition, but instead is a composite deposit that has undergone at least two periods of deposition. The earlier major period of deposition postdates the recession of Bull Lake ice from Horse Butte (Alden, 1953; Witkind, 1969). Hydration rinds from the 80 to 100 cm depth (Fig. 9) display the youngest peak at about 13 microns which indicates that this level largely dates from Bull Lake time (Fig. 21). However, hydration rinds on gravel collected at depths from 2 to 12 m below the surface of the OSP in other locations date from Pinedale time (Pierce, 1979). This indicates that the depositional and erosional history of the OSP is complex and that there may be large paleo-channels in the OSP that were eroded in pre-Pinedale time and filled during Pinedale aggradation.

Only about 80 cm of sediment was aggraded or reworked at the location of sampling pit TO during Pinedale time. As mentioned above, deposition could have been in the form of shallow aggrading streams (Richmond, 1964) or from glacial outburst floods (Pierce, 1979). From evidence presented below, advancing and full Pinedale ice conditions are characterized by aggradation whereas Pinedale recession is characterized by erosion and terrace formation. It must be emphasized that this multi-age OSP interpretation is based on one site only but is compatible with evidence presented by Alden (1953), Richmond (1964), Witkind (1969) and Pierce (1979).

It is difficult to reconcile the disparity between aggradation during Bull Lake recession as postulated by Witkind (1969) but downcutting during Pinedale recession. However, the glacial regime of each ice mass was very different. The Bull Lake glacier extended out into the West Yellowstone Basin to the position of Horse Butte whereas Pinedale ice just exited the mouth of Madison Canyon (Fig. 1). The gradient from the mouth of Madison Canyon to Horse Butte begins at about 1° but soon lessens to about 0.1° (Pierce, 1979). The gradient of the surface is very low, but probably does not reflect the original surface that the Bull Lake ice moved across. Richmond (1964) reported that the gravel ranges in depth from 12 to 30 m and rests on lacustrine sediments impounded by the Bull Lake moraines. Depending on the thickness of the lacustrine deposits, the gradient of the surface may have been much steeper.

The glacier dynamics of Bull Lake ice are further differentiated from Pinedale glacier dynamics because the West Yellowstone flow was not emplaced until at least late Bull Lake time (Pierce and others, 1976). Pierce (1976) reported that Bull Lake ice moved into the West Yellowstone Basin from the southeast through a 15 km gap that is now occupied by the West Yellowstone flow. Pinedale ice advanced into the basin from the northeast and originated at the ice divide located in the vicinity of Obsidian Cliff. If the West Yellowstone flow was emplaced while Bull Lake ice was present, it is likely that there were large jokulhlaups in the basin during this time. The West Yellowstone flow has been K-Ar dated at 117 $\pm$ 8 ka and has concave embayments that suggest that it was erupted against ice (Waldrop, 1975; Waldrop and Pierce, 1975; Pierce, 1975; Pierce and others, 1976; Richmond, 1976b; Pierce, 1979). The eruption of lava in contact with the ice could have provided a mechanism for melting and floods in the West Yellowstone Basin which may have contributed to the deposition of the OSP during late Bull Lake time.

Age of Terraces

There is no apparent difference between the position of the youngest peak from Terraces T0 through T4 (Figs. 10-14), indicating that the terraces were cut in a shorter period of time than the technique can discern. Each surface has a large peak (Figs. 10-14) centered around 6.5 microns with a range from 6 to 7 microns. Because the data sets most probably reflect multi-modal distributions, parametric statistics were originally not used to make inferences about the populations. However, ongoing analyses by Rossi (1990, personal comm.) should provide more reliable estimates of the center of the peaks and the standard error of these estimates by means of parametric curve fitting. For the present, the center (mode) of the large peak for each terrace level is the best measure of rind thickness.

The first method used to obtain a date for the terraces was application to the calibration curve from Pierce and others (1976) (Fig. 21) of an hydration value of 6.5 $\pm$ 0.5 microns, which yields an age of 20,000 $\pm$ 3,000 yr. This age agrees with Pierce and others (1976) in that the terraces are younger than full Pinedale moraines (~30 ka) but older than Pinedale deglacial deposits (~12 ka). Nash (1984) used scarp morphology to date the terraces by calibrating the technique against a minimum ^{14}C date of 7,100 $\pm$ 50 yr for Terrace T4. If this date is too young, the dates obtained from the upper scarps will also be too young. Figure 22 shows the ages Nash (1984) derived for the terraces compared to ages for the terraces derived by the first method in this section. As can be seen from this comparison, most of the ages do not overlap because Nash (1984) estimated that the terraces are significantly younger than estimates from this study.

A second method used to obtain an age for the formation of the terraces was to use hydration rate constants calculated for the Obsidian

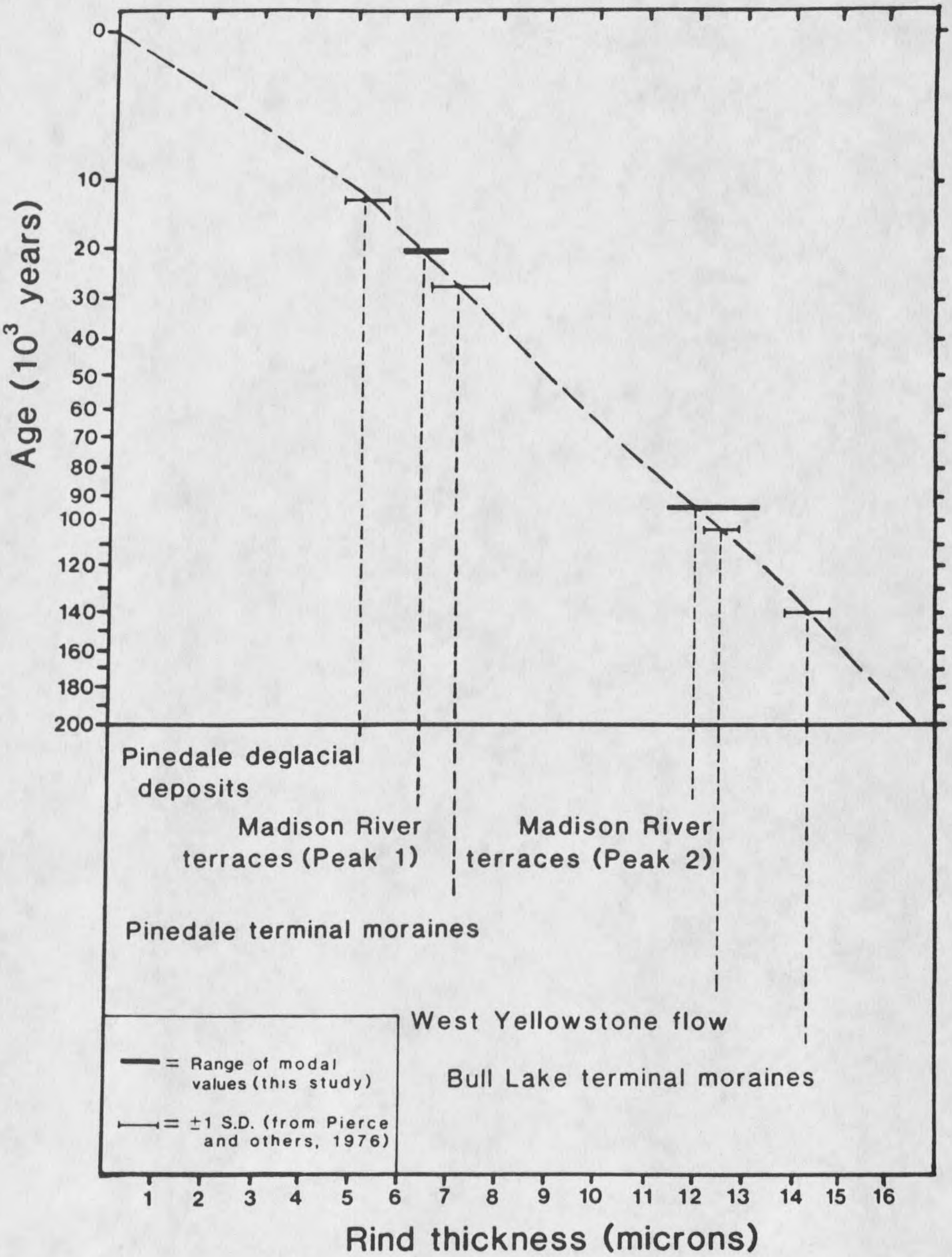


Figure 21 - Hydration calibration curve showing ages of Madison River terraces and an interpreted older event (heavy lines, this study) in relation to glacial deposits in the West Yellowstone area (after Pierce and others, 1976).

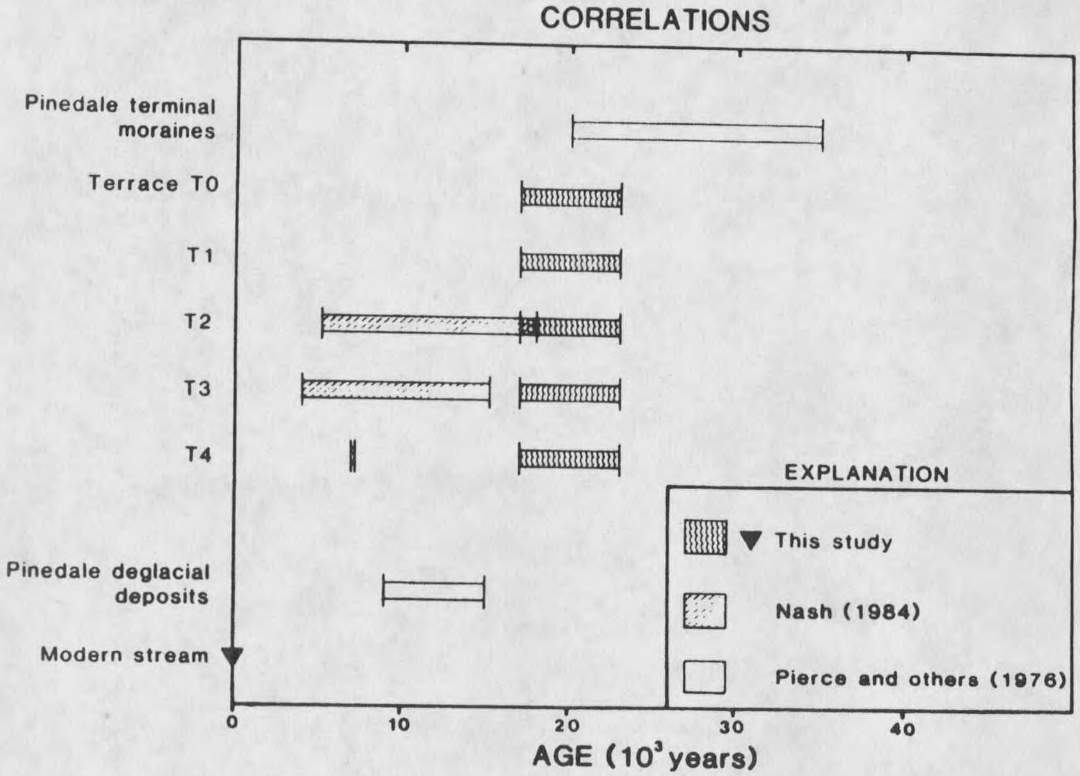


Figure 22 - Correlation chart showing ages of selected surficial geologic features in the West Yellowstone area.

Cliff flow (Michels, 1985) together with an estimated effective hydration temperature (EHT). This method assumes that the obsidian in the OSP is either derived from the Obsidian Cliff flow or hydrates at roughly the same rate. Friedman and Obradovich (1981) examined eight obsidian flows in Yellowstone National Park and found that all of the flows were chemically similar to the Obsidian Cliff flow. Therefore, they postulated that there was no difference in hydration rate among the flows. Pierce and others (1976) reported that most of the material in the OSP was probably derived from the West Yellowstone flow, which was included in the analysis by Friedman and Obradovich (1981).

The Arrhenius equation used by Michels (1985) to relate hydration rate to temperature is the same as that used by Friedman and Long (1976) and listed earlier in this thesis. The working version of this equation for the Obsidian Cliff flow takes the form:

$$k' = 7.05 \exp[77596/8.317\ (1/473.16 - 1/T')]$$

where k' is an unknown (hydration rate at T') and T' is equal to the EHT in K for the site to be dated. To convert the EHT from °C to K use:

$$T_e\ (K) = T_e\ (°C) + 273.16$$

The equation that converts the rate constants to the hydration rate is:

$$k'(365.26)(1000) = \text{rate expressed as microns}^2/1000 \text{ years}$$

Friedman and Obradovich (1981) estimated that the present EHT on the West Yellowstone flow is 3.0 °C and the EHT during glacial times was 0 °C. This is somewhat different than estimates by Pierce and others (1976) of 3.9 °C (present day) and -2.1 °C (glacial). To adjust for the elevation difference between the OSP and the West Yellowstone flow a standard value of 0.5 °C was added to all estimates of EHT. Pierce and others (1976) considered postglacial EHT's to have been in effect from 0 to 11 ka and glacial EHT's to have been in effect from 11 to 75 ka and 127 to 170 ka. The terraces are thought to be post-maximum Pinedale in age (< 20 ka), so glacial EHT's have been operating for roughly half the time and post glacial EHT's have been in effect for the remainder of the time for this site. In this study an EHT of 1.4 °C was chosen because this is an average of present and glacial temperature (+ 0.5 °C for elevation difference) as estimated by Pierce and others (1976) and is thought to represent an integrated estimate of temperatures operating from late Pinedale to the present.

An EHT of 1.4 °C coupled with an hydration rind thickness of 6.5 ± 0.5 microns yields an age of 15,520 +2480 -2300 years for the fluvial terraces. This age is less than that calculated for the terraces by means of the calibration curve (Fig. 21), but is in the same range.

It is hypothesized here that terrace cutting was initiated immediately upon retreat of full Pinedale ice. Rapid terrace formation is not unreasonable in light of work by Thompson and Jones (1986) who determined that rates of incision were associated with the position of an ice front. During times of rapid glacial retreat, incision was also rapid, but during times of slower glacial retreat or still stands, incision was slower or aggradation occurred.

Yellowstone Lake

Reworking Processes

Reworking in the beach environment is largely caused by abrasion. Grogan (1945) studied the shapes of rhyolite gravel on a Lake Superior beach and found that roundness increased away from the source area, while the overall size changed very little. This indicates that fracturing was probably not an active process on this beach, because fracturing would tend to increase the angularity of the gravel while at the same time decreasing the size. Grogan (1945) attributed the increase in rounding to traction transport along the beach face. Sames (1966) studied isotropic (chert and quartzite) pebble associations to determine if littoral and fluvial environments could be distinguished on the basis of roundness and angularity. He found that both environments show a fair degree of roundness, but the fluvial pebble associations had a much higher percentage of angular components and a wider variety of shapes. Dobkins and Folk (1970) also studied the reworking processes of both streams and beaches. They found that, while angularity remained relatively constant during stream transportation, a significant increase in rounding occurred in the beach environment which was also attributed to traction transport on the beach face. From the above studies, it can be concluded that fluvial pebbles are reworked by a combination of abrasion and fracturing, whereas beach pebbles are reworked mainly by abrasion. This conclusion has important ramifications for this study because hydration rinds were measured only on fractured surfaces and consequently reworking processes on the shorelines are not well reflected by the sample measurement technique. However, because abrasion is thought to represent small-scale fracturing, some of the rinds measured on the shorelines could date from the time of beach abrasion.

Age of Shorelines

The rind measurement data for all shorelines (S1-S5) display a prominent peak that ranges from 5.0 to 6.2 microns (Figs. 16-20). At the outset of the study it was hypothesized that each shoreline has a unique temporal signal that is reflected by the thickness of hydration rinds from that particular surface. However, this does not appear to be the case. Instead, all of the shorelines are very similar with regard to hydration rind thicknesses. As with the fluvial terraces, which also had very similar (but thicker) rind thicknesses, it can be hypothesized that the shorelines were formed in a more rapid period of time than the technique can discern. This may be true, but the author believes that the rind signal for all of the shorelines is inherited from some previous geologic event and is not associated with the formation of the shorelines. As with the terraces, a somewhat arbitrary hydration value of 5.5 microns was chosen to represent the shorelines pending further statistical analyses by Rossi (1990, personal comm.).

The shorelines studied at Yellowstone Lake are not older than about 14.5 ka and are probably younger than 12 ka. Lacustrine and alluvial deposits in the Yellowstone Lake Basin have yielded ^{14}C ages of 14,360 $\pm$ 400 (Waddington and Wright, 1974), 14,490 $\pm$ 350 (Richmond, 1976b), 14,130 $\pm$ 370 (Kelley and others, 1978) and 13,650 $\pm$ 650 BP (Richmond, 1976b). Richmond (1976a) estimates that Yellowstone Lake was completely ice free by about 12 ka. At three localities around the lake, Glacier Peak ash has been identified (Richmond, 1976a). This ash has been ^{14}C dated at 11.2 ka in Montana (Mehringer and others, 1984). Richmond (1976a) reported a maximum ^{14}C date of 9060 $\pm$ 300 years associated with an 18–20 m shoreline which he considered to be the first shoreline to form after complete deglaciation of the lake basin. However, because of deformation this shoreline is not everywhere 18–20 m above present lake level. As discussed previously, Meyer and Locke (1986) reported minimum ^{14}C ages of 2495 ± 130 and 1410 ± 160 yrs B.P. for lagoons at 18.4 and 9.1 m, respectively. These minimum and maximum ages indicate that ice-marginal lakes and streams were present in the Yellowstone Lake Basin by about 14.5 ka and that the basin was probably completely ice free by about 12 ka.

Chemical analyses of detrital obsidian grains from the shorelines at Breeze Point (Appendix C) indicate that virtually all of the grains are from the Aster Creek flow (Wayne Hamilton, 1989, written comm.). These grains were randomly selected from different shorelines and from different sample levels. Therefore, it is reasonable to conclude that most of the obsidian gravel on the shorelines at Breeze Point is derived from the Aster Creek flow which is the local bedrock in the area (Fig. 2).

Hydration rate constants for the Aster Creek flow have been determined by Michels (1988). These constants allow numerical age estimates for rind thicknesses if the EHT is known. The working version of the Arrhenius equation relating hydration rate to temperature takes the form of:

$$k' = 5.14 \exp[86834/8.317 \ (1/473.16 - 1/T')]$$

where k' is the hydration rate at T' and T' is equal to the EHT in K for the site. The final equation that gives the hydration rate for the Aster Creek flow is:

k'(365.26)(1000) = rate expressed as microns^2/ 1000 years

Thus if the EHT can be measured or estimated, a numerical date can be generated using the above equation. However, for reasons given below, there is the possibility that the hydration rate constants for the Aster Creek flow are in error.

Friedman and Obradovich (1981) used a glacial soil temperature of −1 °C and directly measured present soil temperature to determine EHT's for various sites in Yellowstone in their study of dating volcanic events using obsidian hydration dating. Friedman and Norton (1981) used the Pallman method to directly measure soil temperature in the Yellowstone area so that present day EHT's could be used in studies employing obsidian hydration dating. Soil temperature was measured at twelve sites around Yellowstone Lake and temperatures along the western shore nearest Breeze Point range from 5.7 to 10.6 °C. Friedman and Norton (1981) hypothesized that the high soil temperatures result from geothermal heat flux. These temperatures were normalized to a site at 7000 feet asl (2134 m) and in direct sunlight. The shorelines at Breeze Point are 7732 feet asl (2357 m) and are shaded because of the dense lodgepole cover. Therefore, the actual present EHT for the study site is probably lower than the values reported by Friedman and Norton (1981) for sites around Yellowstone Lake. Assuming that the present soil temperature represents the high end of the range for EHT and the glacial temperature represents the low end, the actual EHT probably lies between these two end points.

The age of the geologic event that produced the youngest rinds on the shorelines can be estimated by using the Arrhenius equation and the estimated and measured EHT's. Table 5 lists hydration rates and corresponding numerical ages of a 5.5 micron rind using temperatures ranging from -1 °C to 11 °C. Clearly, some of these dates are erroneous given that the Aster Creek flow has been K-Ar dated at 174 ± 4 ka (Richmond, 1986). The EHT for the Breeze Point site must have been near or above 6 °C, if the hydration constants calculated by Michels (1988) are correct. Otherwise, the derived dates are older than the obsidian flow itself. A 5.5 micron rind yields a date of about 48 ka if a temperature of 11 °C (measured at Grant Village) is used. It is unlikely that the EHT has been higher than this value and it is probably much lower, given the near-glacial climate in Yellowstone Park throughout most of the Holocene (Richmond, 1986).

Table 5. Hydration rates and numerical ages for a 5.5 micron rind and effective hydration temperatures ranging from -1 to 11 °C for Aster Creek obsidian (After Michels, 1988).

Temperature	Rate	Age
-1 °C	0.16 microns^2/1000 yrs	1,182,000 yrs
0	0.18	934,000
1	0.21	686,000
2	0.26	525,000
3	0.27	415,000
4	0.31	315,000
5	0.36	233,000
6	0.41	180,000
7	0.47	137,000
8	0.54	104,000
9	0.61	81,000
10	0.70	62,000
11	0.79	48,000

It is possible that the hydration rate constants for the Aster Creek flow (Michels, 1988) are in error. The constants yield rates that are an order of magnitude less than Obsidian Cliff constants for the same temperature, even though the chemistries are very similar (Appendix C). Friedman and Long (1976) reported that increased SiO_2 increases the rate of hydration while increased CaO and MgO decrease hydration rate. Chemical analyses of Aster Creek obsidian (Michels, 1988a, 1988b) show a higher silica content than Obsidian Cliff (Friedman and Long, 1976; Michels, 1985) and virtually identical values of CaO and MgO (Appendix C). These analyses indicate that the Aster Creek flow should hydrate at a rate faster than or equal to that of the Obsidian Cliff flow.

Friedman and Obradovich (1981) analyzed the chemistry of eight different Obsidian flows in Yellowstone Park. The ages of these flows range from 77 to 399 ka. All of the flows were chemically similar to the Obsidian Cliff flow which led Friedman and Obradovich (1981) to conclude that all of the flows had the same hydration rate as that calculated for Obsidian Cliff. The Aster Creek flow was not included in this analysis, but it is similar in both chemistry and age to the Obsidian Cliff flow.

From the above discussion it is reasonable to assume that the hydration rate constants calculated for the Aster Creek flow (Michels, 1988) are in error. Michels (1990, personal comm.) indicated that the most likely source of error in determining hydration rate constants was calculating the activation energy based on linear regression of induced hydration measurements. Induced hydration is accomplished at high (150-250 °C) temperatures so the extrapolation to ambient temperatures required by this technique could introduce significant error in determining the activation energy of a particular glass. Michels (1990, personal comm.) stated that he requires a correlation coefficient of at least 0.9997 in order to accept the derived activation energy value. The correlation coefficient for determining the activation energy for the Aster Creek flow is 0.9997 (Michels, 1988) which indicates that this value just met the minimum requirements as defined by Michels (1990, personal comm.). An error of just 10 % in determining the activation energy can change the calculated hydration rates by an order of magnitude. This potential source of error could account for the large difference in hydration rate constants between the Aster Creek and Obsidian Cliff flows. Assuming that the Aster Creek hydration rate constants are in error, the constants calculated for the Obsidian Cliff flow (Michels, 1985) will be used to estimate ages for the shorelines at Breeze Point.

An EHT of 1.0 °C was used for the shorelines at Breeze Point because this temperature is thought to approximate an integrated average between glacial temperatures (-1 °C) and present soil temperatures. Less weight was given to present soil temperatures because a cooler climate has prevailed throughout most of the Holocene (Richmond, 1986).

A rind thickness of 5.5 ± 0.5 microns coupled with an EHT of 1.0 °C yields an age of 12,270 +2330 -2130 years for the shorelines at Breeze Point. This age agrees well with estimates of deglaciation for the Yellowstone Lake Basin (Richmond, 1976a; 1976b; 1986) but is thought to be too old to be associated with the formation of the shorelines. Richmond (1976a) reported ^{14}C dates for the shorelines formed after deglaciation of the basin ranging from a maximum age of 9060 ± 300 yrs B.P. for Shoreline S7(?) (18-20 m) to a minimum age of 620 ± 250 yrs B.P. for

Shoreline S2(?) (3 m). Meyer and Locke (1986) reported minimum ^{14}C ages for Shorelines S6 and S5 (Shorelines S5 and S4-this study) of 2495 $\pm$ 130 and 1410 $\pm$ 160 yrs B.P., respectively. Locke (1989) also reported two minimum ^{14}C ages of about 6 ka for Shoreline S6 (Shoreline S5-this study). The hydration age displayed by the shorelines at Breeze Point is considerably older than any of the maximum and minimum ^{14}C ages associated with Holocene shorelines in the lake basin. Therefore, the rinds must be associated with an older resetting event or the EHT must be higher.

It is hypothesized that the rinds from the shorelines effectively date the deglaciation of the Yellowstone Basin. A rind thickness of 5.5 microns closely matches rind thicknesses from deglacial deposits in the Obsidian Cliff area (Fig. 21). Pierce and others (1976) also estimated that deglaciation of the Yellowstone Plateau occurred about 12 ka. This hypothesis is further supported by the lack of an effective reworking mechanism for the beach environment. Consequently, the rinds would be expected to date from the last effective reworking mechanism which was the waning Pinedale ice mass.

Estimates can be made as to how thick hydration rinds dating from the occupation of the shorelines would be if coastal processes were an effective reworking mechanism. Assuming an age of from 2 to 9 ka for the shorelines and an EHT of 1 ^{o}C the hydration rind values should range from about 2.2 to 4.7 microns. All of the shorelines have hydration values in this range (Appendix F). These rinds may date from the occupation of a given shoreline but, unfortunately the nonparametric density estimation techniques as used in this study cannot determine if these thin rinds form a discrete cluster of points. For the sake of argument, estimates of the age of the shorelines can be made if these rinds are assumed to have formed by coastal reworking. Using an EHT of 1 ^{o}C, hydration rate constants for the Obsidian Cliff flow (Michels, 1985) and minimum rind thicknesses from 2.74 to 3.77 (Appendix F), the shorelines yield minimum ages of 3050 to 5770 years. These ages are in reasonable agreement with some of the minimum ^{14}C ages as discussed above.

The shorelines at Yellowstone Lake must have formed after 12 ka but unfortunately the techniques used in this study cannot further resolve the ages of these features with any degree of certainty. It is thought that larger sample sizes of lakeshore gravel would do little to assist in the dating of these features because of the relative ineffectiveness of the beach environment to produce fresh-fractured surfaces.

Interpretation of Older Events

Aggregate data from Terrace T0 (Fig. 10) forms two distinct peaks, in addition to the one centered at about 6.5 microns, that represent events that occurred before late Pinedale time. These peaks are centered at about 12 and 22 microns, respectively. The peak at about 12 microns fits reasonably well with hydration rinds attributed to Bull Lake glaciation by Pierce and others (1976) (Fig. 21). They measured a total of 64 hydration rinds spread over five sites but in this study a total of 158 measurements were taken at one site. Of the 158 measurements, only about 21 % or 33 rinds fell under the peak centered at about 12 microns (Richard Rossi, 1990, personal comm.). Consequently, because of the relatively small sample sizes, it is difficult to determine if these two groups of rinds came from the same population, namely Bull Lake reworked glacial pebbles.

Another hypothesis is that the group of rinds centered around 12 microns date from the emplacement of the West Yellowstone flow. Pierce and others (1976) measured just six rinds interpreted as forming along original cooling cracks on the West Yellowstone flow, but they are all clustered around 12.5 microns, which is a close match to the 12 micron peak from this study. This hypothesis is further supported by Pierce (1979) who stated that most of the material in the OSP was probably derived from the West Yellowstone flow.

The peak centered at about 22 microns (Fig. 10) does not seem to fit into the hydration-event model developed by Pierce and others (1976). The thickest rinds measured in that study were only about 17 microns. Pierce and others (1976) attributed the thickest rinds to cracks formed during the emplacement of the Obsidian Cliff flow. Again, only about six rinds were measured to support this claim. It is very possible that rinds clustered around 22 microns date original cooling cracks and that the rinds measured by Pierce and others (1976) are anomolously thin. This is supported by the fact that a substantial number of rinds (™20) make up the peak centered at 22 microns, which indicates that there probably was some event associated with this peak. And obviously, these rinds cannot date from before the emplacement of the Obsidian Cliff flow (assuming that all material in the OSP has the same range of hydration rate constants as the Obsidian Cliff flow). Assuming that the rinds grouped around 22 microns date from the original emplacement of the Obsidian Cliff flow, a recalibration of the hydration curve from Pierce and others (1976) (Fig. 21) might be warranted. Recalibrating this curve may have the effect of changing the dates for the Bull Lake Glaciation.

An hypothesis that integrates data from Pierce and others (1976) and from this study is that the peak at about 12 microns dates from the Eowisconsin (Richmond, 1986), rinds from 17 to 18 microns date from Bull Lake time and rinds at about 22 microns date from original cooling cracks. The Eowisconsin glaciation occurred from about 80 to 117 ka (Richmond, 1986) and is thought to be less extensive than both the Bull Lake and Pinedale Glaciations. Even though the aggregate data from Terrace TO (Fig. 10) does not show a peak at 17 to 18 microns, many of the density diagrams for both the terraces and shorelines (Figs. 11-20) show a concentration of rind thicknesses in this range.

It is not known why measurement data thicker than about 6.5 microns for the terraces and about 5.5 microns for the shorelines does not seem to consistantly cluster around one or several points. It is likely that reworking during the most recent event removed most of the existing rinds, thereby leaving only a fragmentary record of older reworking events. The spread caused by natural variation in hydration rates would be expected to be larger for older events which has the effect of deemphasizing data concentrations. The lack of clustering may also reflect a complex reworking history that is beyond the scope of this study to decipher.

The Obsidian Hydration Technique

Obsidian hydration dating was originally developed to date archaeological artifacts and has since been adapted to date geologic events and deposits. In all studies known to the author, a single set of hydration rate constants has been used to calculate ages for artifacts or geologic events if the obsidian is thought to be derived from a single source. In some cases, when multiple sources are likely, a single rate is still applied (e.g. Friedman and Obradovich, 1981). The practice of using a single group of hydration rate constants for a given rind or group of rinds has the effect of ignoring natural variation within obsidian flows which may be the cause of scatter in the rind thicknesses.

The scatter of rind thicknesses around each of the peaks for both the terraces (Figs. 10-14) and the shorelines (Figs. 16-20) is attributed to a number of sources. The first, and probably most important, is varying chemistry between and within obsidian flows. Hydration is a function of both temperature and chemistry of a particular rhyolitic glass. As discussed above, Friedman and Long (1976) reported that the amount of SiO_2, CaO and MgO affected the rate of hydration. The amounts of these three constituents cannot be expected to be identical for different obsidian flows. Friedman and Obradovich (1981) reported differences in hydration rates for many flows around the western United States. They also analyzed eight different flows in Yellowstone Park and found that all had similar chemistries. However, the analyses for the different flows in Yellowstone were not reported, and consequently it cannot be determined just how similar these flows are with respect to their SiO_2, CaO and MgO content.

The apparent variability in chemistry within an individual flow probably results from the technique used to analyze the glass in addition to natural variation. Appendix C (Table 7) shows atomic absorption spectroscopy analyses for multiple samples identified as Aster Creek obsidian (Michels, 1988b). As can be seen, considerable variability exists in the amounts of the chemical constituents. In particular, the amounts of CaO and MgO vary by as much as 30 and 50 %, respectively. Appendix C (Table 8) lists two separate analyses of the Obsidian Cliff flow, which also shows considerable variability in key elements. Some of this variation is presumed to be inherent in the glass, but a certain percentage probably results from the analytic technique. Unfortunately, Friedman and Long (1976) and Michels (1985; 1988b) did not report the errors associated with each of the chemical analyses.

Variation in key elements within a single flow indicates that hydration rate should also vary within a single flow, thereby casting doubt on the practice of generating a single group of hydration rate constants for an entire obsidian flow. There is probably a range of hydration rate constants for each individual obsidian flow that is dictated by the variation in chemistry for that flow. Table 6 shows different hydration rates for the Obsidian Cliff flow calculated for the same temperatures by Friedman and Long (1976) and Michels (1985). The disparity between these rates indicates that there is variation in hydration rate within a given flow. As an example, if an age is calculated using constants supplied by Friedman and Long (1976) and Michels (1985) and assuming a rind thickness of 6.5 microns and an EHT of 6.0

°C, there is a difference in age of about 14 %. This translates into an apparent age difference of about 700 years using these parameters.

The variation of hydration rate due to chemical differences within individual flows has important ramifications for archaeologists and geologists who use obsidian hydration dating techniques. In the past, generally only one set of rate constants was generated for an individual flow (Michels, 1985, 1988). These constants are periodically revised (Joseph Michels, 1990, personal comm.), but the underlying assumption is that each obsidian flow possesses unique hydration rate constants. As can be seen from the above example, the assumption of a single rate for individual flows is not justified and may lead to erroneous conclusions about the age of a particular artifact, archaeological site or geologic reworking event.

Table 6. Hydration rates and corresponding ages calculated for a 6.5 micron rind for Obsidian Cliff obsidian using the same temperatures.

Temperature (°C)	Rate (microns^2/1000 yrs)		Age	
	1	2	1	2
6.0	3.10	2.88	4396	5094
6.9	3.40	3.21	3654	4100
19.7	16.00	13.76	165	223
22.7	22.60	19.00	83	117

after (1) Friedman and Long (1976) and (2) Michels (1985)

Friedman and Obradovich (1981) analyzed eight obsidian flows in the Yellowstone region and found that all had similar chemistries. Therefore, they postulated that there was no variation in hydration rates between these flows. Again, the analyses of these individual flows was not reported which makes it quite difficult to assess the similarity of the flows. In light of the fact that individual flows possess considerable variability, it does not seem likely that eight different flows would possess identical hydration rate constants.

Michels and Bebrich (1971) plotted a histograph using 233 rind measurements which range from about 1 to 5 microns from an archaeological site in Mexico (Fig. 23). This range in rind thickness represents a period of time from 3584 B.C. to 1744 A.D. Most of the measurements fall between 2 and 3 microns which corresponds to an age range from 34 B.C. to 1077 A.D., but there are many subsidiary peaks. The spread associated with these measurements was attributed to "an unusually long and complex history of occupation". From the above discussion, it is possible that much of the spread is, instead, caused by differing hydration rates for many of the artifacts. This might imply that the occupation history of this site was not as long and complex as Michels and Bebrich (1971) concluded.

As can be seen from the above examples, workers using obsidian hydration dating techniques must recognize that hydration rate constants can vary between and within individual obsidian flows. Therefore, the limitations of using a single rate (or a single rind) for an archaeological site or geologic reworking event must be recognized and accounted for. The ideal method would be to obtain hydration rate constants for each artifact or sample dated. However, because obtaining hydration rate constants is a destructive and expensive procedure, it is not always possible due to financial limitations and is usually not desirable due to the cultural value of obsidian artifacts.

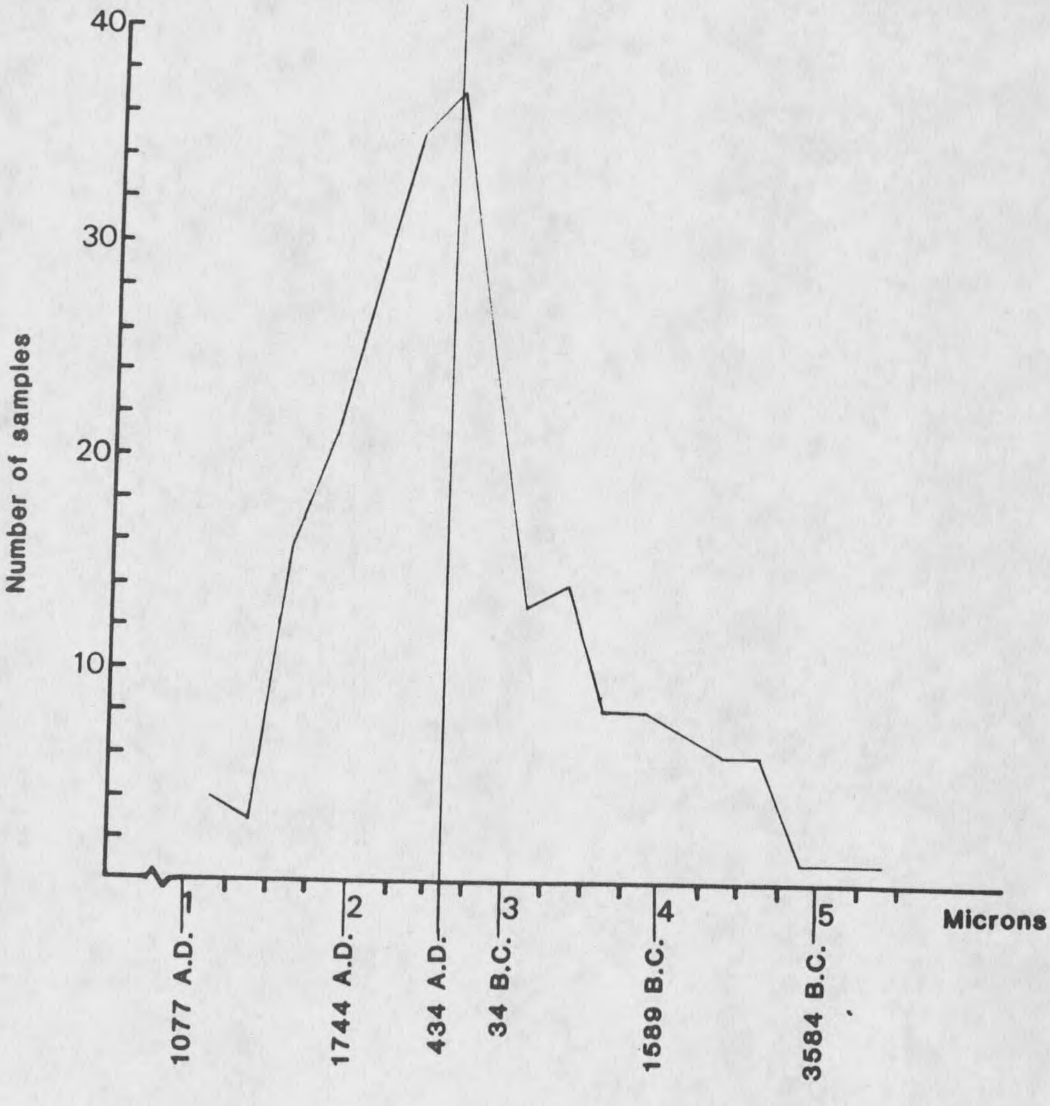


Figure 23 - Histograph of 233 rind measurements from an archaeological site in Mexico (after Michels and Bebrich, 1971).

Another source of scatter for rind measurements is the standard measurement error. The measurement error for this study was determined to be 0.3 microns. This value is significant with regard to individual rind measurements, but is probably negligible in comparison to scatter introduced by differing hydration rates as discussed above. Other sources of scatter include differences in time of reworking and fracturing not related to the event of interest. This fracturing may take the form of bioturbation (root throw or burrowing) or sampling technique and preparation.

The uncertainty introduced by rind measurement scatter for artifacts and geologic events can be minimized by taking multiple rind measurements. To characterize a geologic site or horizon, where obsidian has been naturally worked, a minimum of about 150 measurements must be taken (Richard Rossi, 1990, personal comm.). This number is required because oftentimes obsidian has gone through several reworking events and the number of rind thicknesses is distributed among several modes. The distribution has the effect of lowering the number of rinds used to characterize an individual event. Obsidian artifacts have readily identifiable, event-specific rinds that can be used to characterize an archaeological site. Therefore, considerably less rind measurements are needed to determine a date for an occupation site or horizon.

CONCLUSIONS

The original purpose of this study was to develop an obsidian hydration dating technique that could be used to date Quaternary deposits that contained detrital obsidian. Two types of field settings were investigated. The first was a group of fluvial terraces cut by the ancestral Madison River near West Yellowstone, MT in post-Pinedale time. Hydration rinds were examined from the modern stream in order to determine if it serves as an appropriate analog for the stream that cut the terraces. It was found that the modern stream possesses significantly more zero thickness rinds than the terraces which leads to the conclusion that reworking does occur in the fluvial environment. Therefore, the terraces should possess unique and characteristic hydration rind thicknesses that can be used to determine their ages. Because rinds of several different thicknesses on individual grains were also found in the modern stream, only partial reworking occurs in this environment. This indicates that each naturally worked obsidian pebble can potentially record several reworking events.

Examination of pebbles sampled at 20 cm increments from the Obsidian Sand Plain (Terrace T0) indicates that solar heating did not appear to affect the rate of hydration in the upper few decimeters at this site. The aggregate data from Terrace T0 also indicates that the Obsidian Sand Plain might date from several periods of deposition. All of the sampling levels except for the 80-100 cm level possess a strong rind signal at about 6.5 microns. The lower level has strong peaks at about 13 and 22 microns which probably date from Bull Lake reworking and original cooling cracks associated with the emplacement of an obsidian flow. Therefore it is concluded that the Obsidian Sand Plain dates from at least two periods of deposition at this site.

Terraces T1 through T4 have a prominent rind signal at about 6.5 microns, indicating that they were cut in a shorter period of time than the technique can discern. Two methods were used to obtain an age for the terraces. The first was to use a rind value of 6.5 $\pm$ 0.5 microns and the calibration curve of Pierce and others (1976) which yields an age of 20,000 $\pm$ 3000 years. The second method was to use the same rind value coupled with hydration rate constants for the Obsidian Cliff flow calculated by Michels (1985) and an effective hydration temperature of 1.4 °C. The second method yields an age of 15,520 + 2480 −2300 for the terraces. Both of these ages are older than those calculated by Nash (1984) for the terraces. However, if the calibration ^{14}C date of 7,100 $\pm$ 50 yrs B.P. used by Nash (1984) is too young, his terrace ages are also too young. It is hypothesized that terrace formation was accomplished immediately upon retreat of maximum Pinedale ice from the mouth of Madison Canyon.

The second field setting involved raised shorelines of post-glacial Yellowstone Lake. The shorelines also all have similar rind thicknesses to one another. A rind value of 5.5 microns coupled with hydration rate constants for the Obsidian Cliff flow (Michels, 1985) and an effective hydration temperature of 1.0 °C yields an age of 12,270 +2330 −2130 for the lakeshores. Considering that all dated post-glacial shorelines in the Yellowstone Lake basin are younger than about 9 ka, it is concluded that the rind signal associated with the shorelines dates a previous reworking event. It is hypothesized that the rinds on the shorelines date from latest Pinedale reworking.

Reworking processes in the fluvial environment differ from those in the beach environment in that the dominant process in streams is fracture whereas the dominant process on beaches is abrasion. In this study rinds were measured only on fractured surfaces because rinds on abraded edges were discontinuous and of varying thickness. The technique of measuring rinds only on fractured surfaces thereby excluded most rinds that were reworked in the beach environment.

Hydration rate is dependent on temperature and the chemistry of the glass. Because chemistry varies within and between obsidian flows, rate would also be expected to vary. Therefore, it is not necessarily appropriate to use a single group of hydration rate constants determined from a single sample to calculate ages for artifacts and geological samples without recognizing this limitation. However, the cost of calculating hydration rate constants for every sample to be dated is prohibitively expensive.

The obsidian hydration dating technique developed in this study is very appropriate for dating fluvial reworking events but does not appear to be applicable to dating material reworked in the lacustrine environment. This technique can also be followed, with minor adaptations, to date glacial features as has been done by Pierce and others (1976). This dating technique should serve as a useful tool to other Quaternary workers desiring to unravel the events of the past.

REFERENCES CITED

REFERENCES CITED

Alden, W.C., Physiography and glacial geology of western Montana and adjacent areas: U.S.G.S. Prof. Paper 174, 133 p.

Baily, A.L., 1984, Submerged strands of Yellowstone Lake as indicators of uplift: in Yellowstone Physical Science Symposium, 2nd, Abstracts with Program: Yellowstone National Park, Wyoming, National Park Service, p. 5.

Baker, R.G., 1976, Late Quaternary vegetational history of the Yellowstone lake basin, Wyoming: U.S. Geological Survey Prof. Paper 729-E, 48 p.

Birkeland, P.W., 1984, Soils and geomorphology: Oxford University Press, New York, 372 p.

Blackwelder, E., 1915, Post-cretaceous history of the mountains in central western Wyoming: Jour. of Geology, v. 23, no. 4, P. 97-117, 193-217, 307-340.

Clark, D. L., 1964, Archaeological chronology in California and the obsidian hydration method: Annual report, Archaeological Survey, 1963-64, University of California, Los Angeles, p.139-230.

Christiansen, R.L., 1984, Yellowstone magmatic evolution: its bearing on understanding large-volume explosive volcanism: in Explosive Volcanism: Inception, evolution, and hazards: National Research Council, Geophysics Study Committee, p. 84-95.

Christiansen, R.L., and Blank, H.R., 1972, Volcanic stratigraphy of the Quaternary rhyolite plateau in Yellowstone National Park: U.S.G.S. Prof. Paper 729-B, 18 p.

Church, M.A., 1972, The Baffin Island sandurs: a study of Arctic fluvial processes: Geol. Survey Canada Bull., 216, 208 p.

Dobkins, J.E., and Folk, R.L., 1970, Shape development on Tahiti-Nui: Jour. of Sedimentary Petrology, v. 40, no. 4, p. 1167-1203.

Dyson, J., 1960, Precise measurement by image splitting: Journal of the Optical Society of America, v. 50, no. 8, p. 754-757.

Friedman, I, 1968, Hydration rind dates rhyolite flows: Science, v. 159, p. 878-879.

Friedman, I., and Smith, R.L., 1960, A new dating technique using obsidian. Part 1, The development of the method: American Antiquity, v. 25, p. 476-493.

Friedman, I., and Long, W., 1976, Hydration rate of obsidian: Science, v. 191, p. 347-352.

Friedman, I., and Norton, D.L., 1981, Ground temperature measurements: Part III, Ground temperatures in and near Yellowstone National Park: USGS Prof. Paper 1203, p. 23-39.

Friedman, I., and Obradovich, J., 1981, Obsidian hydration dating of volcanic events: Quaternary Research, v. 16, p. 37-47.

Golden Software, 1984, Golden Graphics System, Surf: Golden, CO, 22 p.

Grogan, R.M., 1945, Shape variation of some Lake Superior beach pebbles: Jour. of Sedimentary Petrology, v. 15, p. 3-10.

Hall, R.D., and Heiny, J.S, 1983, Glacial and postglacial stratigraphy and chronology, north Willow Creek and Cataract Creek drainage basins, Eastern Tobacco Root Range, southwestern Montana, U.S.A.: Arctic and Alpine Research, v. 15, no. 1, p. 19-52.

Hamilton, W.L., 1985, Analysis of submerged terraces, Yellowstone Lake: in Yellowstone Physical Science Symposium, 3rd, Abstracts with Program: Yellowstone National Park, Wyoming, National Park Service, p. 4.

Kelley, L., Spiker, E., and Rubin, M., 1978, Geological Survey, Reston, Virginia, Radiocarbon Dates XIV: Radiocarbon, v. 20, no. 2, p. 283-312.

Komar, P.D., 1976, Beach processes and sedimentation: Prentice-Hall, Inc., Englewood Cliffs, NJ, 429 p.

Leach, B.F., 1977, New perspectives on dating obsidian artefacts in New Zealand: New Zealand Journal of Science, v. 20, p. 123-138.

Locke, W.W., 1979, Etching of hornblende grains in arctic soils: an indicator of relative age and paleoclimate: Quaternary Research, v. 11, p. 197-212.

_________, 1989, Postglacial vertical displacements in the Yellowstone Caldera: GSA abstracts with programs, v. 21, no. 5, march 1989, p. 108.

Locke, W.W., and Meyer, G.A., 1985, Late Holocene vertical deformation, Yellowstone National Park, Wyoming: G.S.A. Abstracts with Programs, v. 17, p. 645.

Lundstrem, S.C., 1986, Soil stratigraphy and scarp morphology studies applied to the Quaternary geology of the southern Madison Valley, Montana: unpublished Master's thesis, Humboldt State University, Arcata, CA, 53 p.

Mehringer, P.J., Jr., Sheppard, J.C., and Froit, F.F., Jr., 1984, The age of Glacier Peak tephra in west-central Montana: Quaternary Research, v. 21, no. 1, p. 36-41.

Meighan, C.W., 1983, Obsidian dating in California: theory and practice: American Antiquity, v. 48, no. 3, p. 600-609.

Meyer, G.A., 1986, Genesis and deformation of Holocene shoreline terraces, Yellowstone Lake, Wyoming [M.S. Thesis]: Bozeman, Montana, Montana State University, 94 p.

Meyer, G.A., and Locke, W.W., 1986, Origin and deformation of Holocene shoreline terraces, Yellowstone Lake, Wyoming: Geology, v. 14, p. 699-702.

Michels, J.W., 1983, Hydration rate constants for Camas-Dry Creek obsidian, Clark County, Idaho: Mohlab Technical Report no. 26, State College, PA, 4 p.

_________, 1985, Hydration rate constants for Obsidian Cliff Obsidian, Yellowstone National Park, Wyoming: Mohlab Technical Report no. 2, State College, PA, 4 p.

_________, 1988a, Hydration rate constants for Aster Creek flow obsidian: Mohlab Technical Report no. 84, State College, PA, 4 p.

_________, 1988b, Written communication: Dept. of Anthropology, State College, PA.

_________, 1990, Personal communication: Dept. of Anthropology, State College, PA.

Michels, J.W., and Bebrich, C.A., 1971, Obsidian hydration dating, in Michael, H.N., and Ralph, E.K. eds., Dating Techniques for the Archaeologist: The MIT Press, Cambridge, Mass., p. 164-221.

Nash, D.B., 1984, Morphologic dating of fluvial terrace scarps and fault scarps near West Yellowstone, MT: GSA Bulletin, v. 95, p. 1413-1424.

Parker, M.L., Jozsa, L.A., Johnson, S.G. and Bramhall, P.A., 1984, Tree-ring dating in Canada and the northwestern U.S.: in Mahaney, W.C., ed., Quaternary Dating Methods: Elsevier Publishing Company, NY, NY, p. 211-226.

Pelton, J.R., and Smith, R.B., 1982, Contemporary vertical surface displacements in Yellowstone National Park: Jour. of Geophysical Research, v. 87, p. 2745-2761.

Pierce, K.L., 1979, History and dynamics of glaciation in the northern Yellowstone National Park area: USGS Prof. Paper 729-F, 90 p.

Pierce, K.L., Obradovich, J.D., and Friedman, I., 1976, Obsidian hydration correlation and dating of Bull Lake and Pinedale Glaciations near West Yellowstone, Montana: GSA Bull., v. 87, p. 703-710.

Richmond, G.M., 1964, Glacial geology of the West Yellowstone basin and adjacent parts of Yellowstone National Park: U.S.G.S. Prof. Paper 435-T, p. 223-236.

__________, 1974, Surficial geologic map of the Frank Island quadrangle, Yellowstone National Park, Wyoming: U.S. Geological Survey, Misc. Investigations Map I-642.

__________, 1976a, Surficial geologic history of the Canyon Village quadrangle, Yellowstone National Park, Wyoming, for use with Map I-652: U.S. Geological Survey, Bulletin 1427, 35 p.

__________, 1976b, Pleistocene stratigraphy and chronology in the mountains of western Wyoming:in Mahaney, W.C., (ed.), Quaternary Stratigraphy of North America, Dowden, Hutchinson and Ross, Stroudsburg, PA, p. 353-379.

Ross, C.S., and Smith, R.L., 1955, Water and volatiles in volcanic glasses: American Mineralologist, v. 40, p. 1071-1089.

Rossi, R., 1989, Personal communication: Dept. of Statistics, Montana State University, Bozeman, MT.

Sames, C.W., 1966, Morphometric data of some recent pebbles associations and their application to ancient deposits: Jour. of Sedimentary Petrology, v. 36, p. 126-142.

Shackleton, N.J., and Opdyke, N.D., 1973, Oxygen isotope and paleomagnetic stratigraphy of equatorial pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 year and 10^6 year scale: Quaternary Research, v. 3, p. 39-55.

Tarter, M.E., and Kronmal, R.A., 1976, An introduction to the implementation and theory of nonparametric density estimation: The American Statistician, v. 30, no. 3, p. 105-112.

Terasmae, J., 1984, Radiocarbon dating: some problems and potential developments: in Mahaney, W.C., ed., Quaternary Dating Methods: Elsevier Publishing Company, NY, NY, p. 1-16.

Thiel, G.A., 1940, The relative resistance to abrasion of mineral grains of sand size: Jour. of Sedimentary Petrology, v. 10, p. 103-124.

Thompson, A., and Jones, A, 1986, Rates and causes of proglacial river terrace formation in southeast Iceland: an application of licheno-metric dating techniques: Boreas, v. 15, p. 231-246.

Twenhofel, W.H., 1945, The rounding of sand grains: Jour. of Sedimentary Petrology, v. 15, p. 59-71.

Waddington, J.C.B., and Wright, H.E., Jr., 1974, Late Quaternary vegetation changes on the east side of Yellowstone Park, Wyoming: Quaternary Research, v. 4, p. 383-446.

Waldrop, H.A., 1975, Surficial geologic map of the West Yellowstone quadrangle, Yellowstone National Park and adjoining area, Montana, Wyoming and Idaho: U.S. Geol. Survey, Miscellaneous Geologic Investigations Map I-648.

Waldrop, H.A., and Pierce, K.L., 1975, Surficial geologic of the Madison Junction quadrangle, Yellowstone National Park, Wyoming: U.S. Geol. Survey, Miscellaneous Geologic Investigations Map I-651

Wehmiller, J.F., 1984, Relative and absolute dating of Quaternary mollusks with amino acid racemization: evaluation, applications, and questions: in Mahaney, W.C., ed., Quaternary Dating Methods: Elsevier Publishing Company, NY, NY, p. 171-194.

Witkind, I.J., 1969, Geology of the Teepee Creek quadrangle, Montana-Wyoming: U.S.G.S. Prof. Paper 609, 101 p.

Ziegler, V., 1911, Factors influencing the rounding of sand grains: Jour. of Geology , v. 19, p. 645-654.

APPENDICES

APPENDIX A

SAMPLE PREPARATION FLOW CHART

Figure 24 - Flow chart for sample preparation procedures.

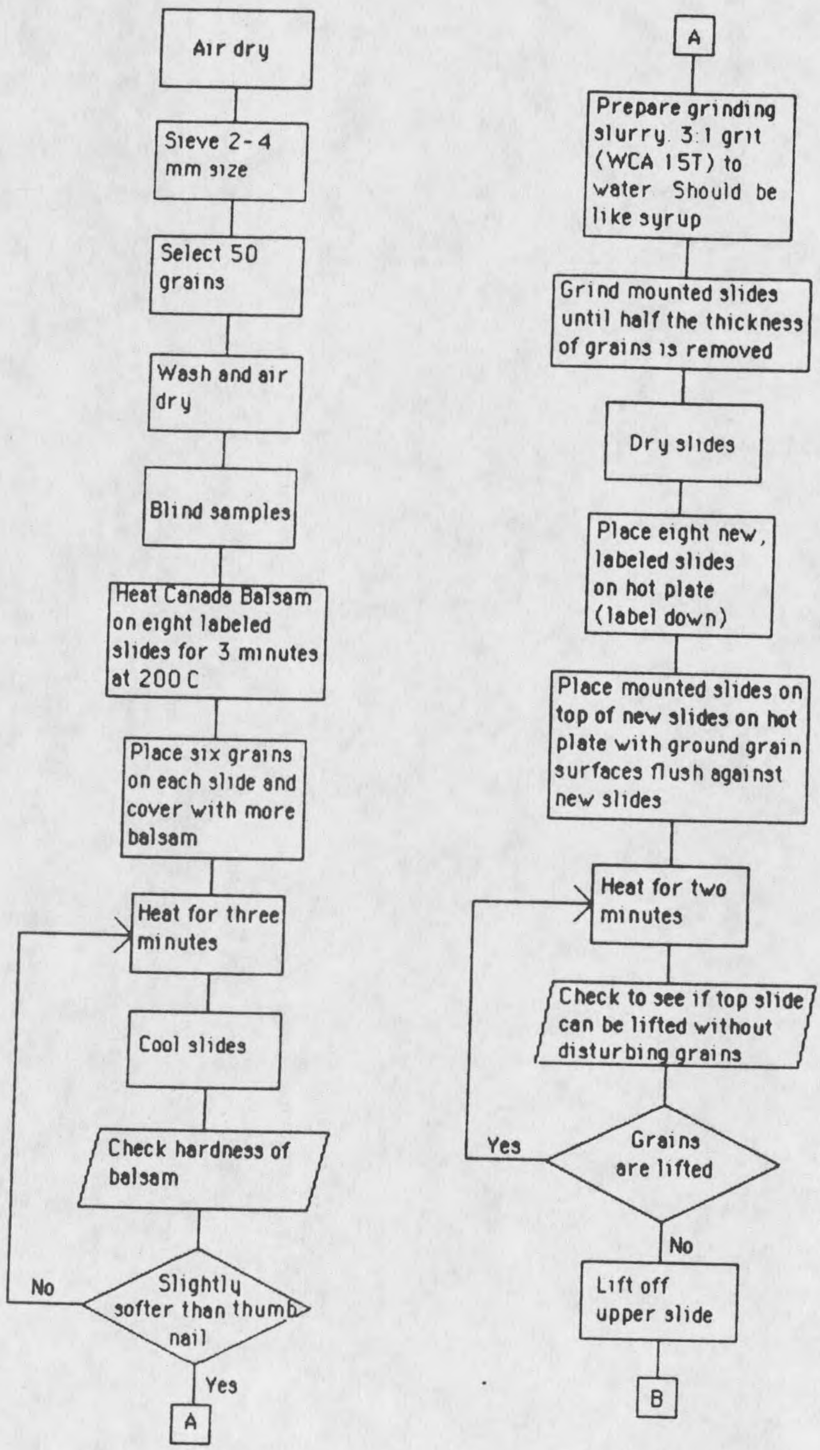
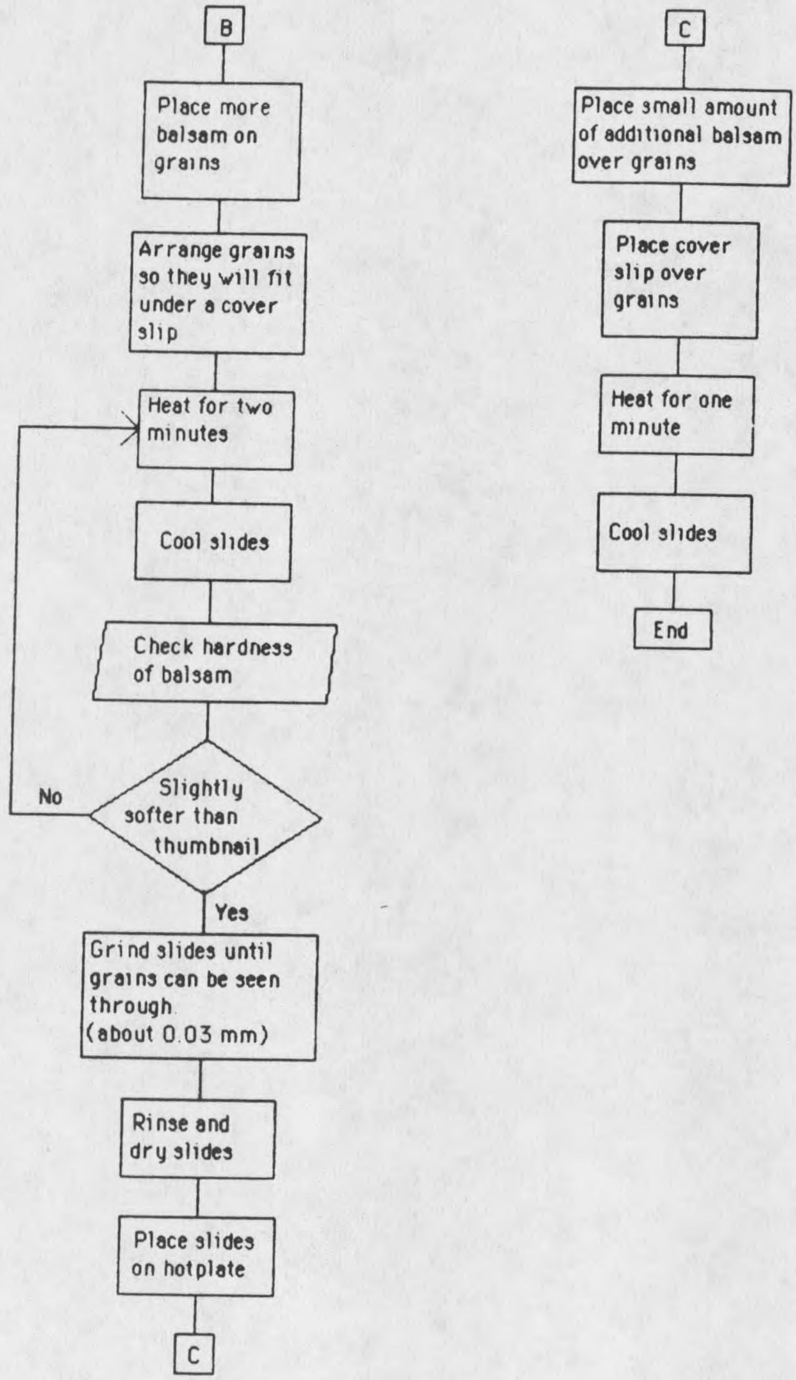


Figure 24 - (continued).



APPENDIX B

HYDRATION RIND MEASUREMENT FLOW CHART

Figure 25 - Flow chart for hydration rind measurement.

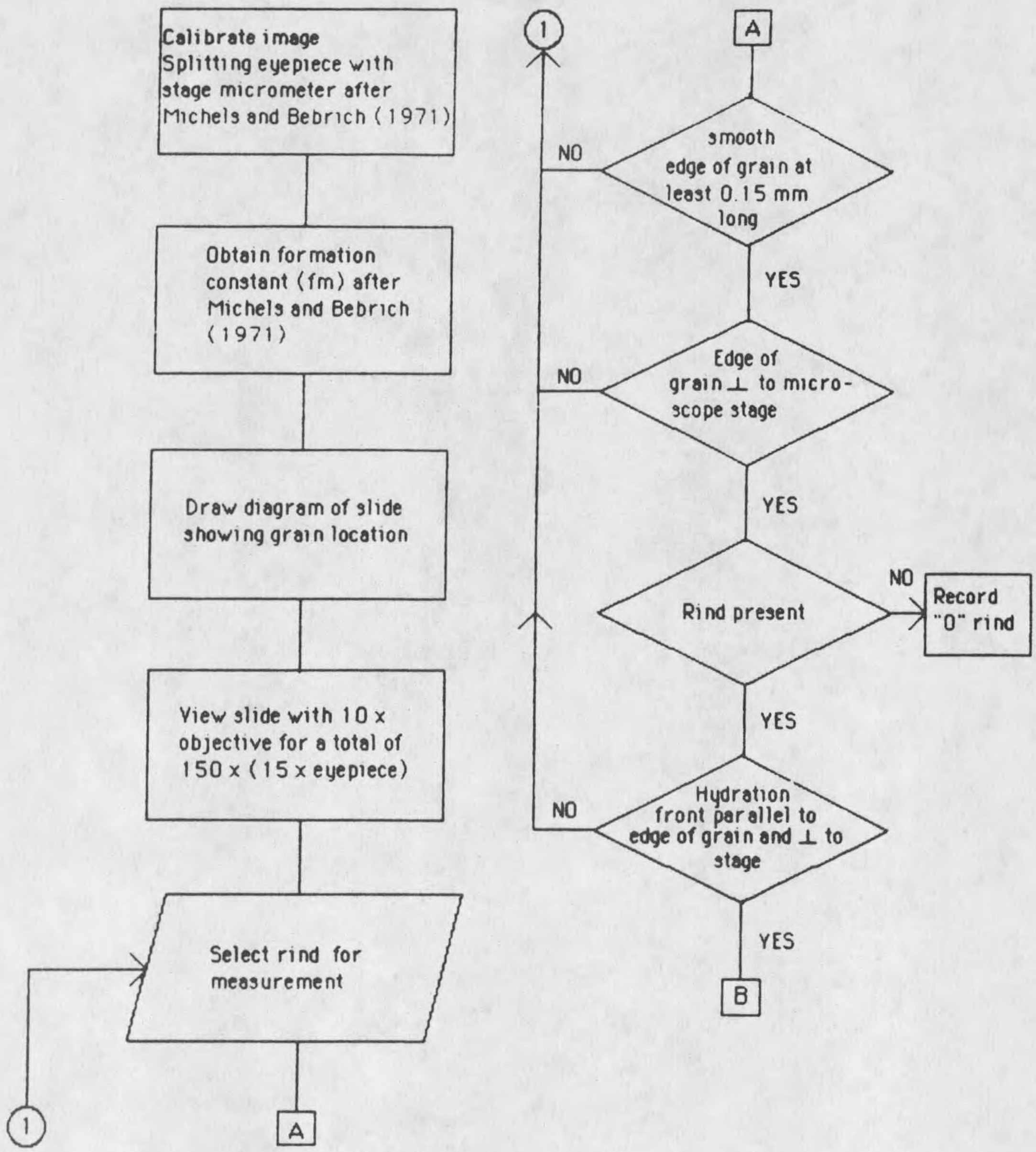
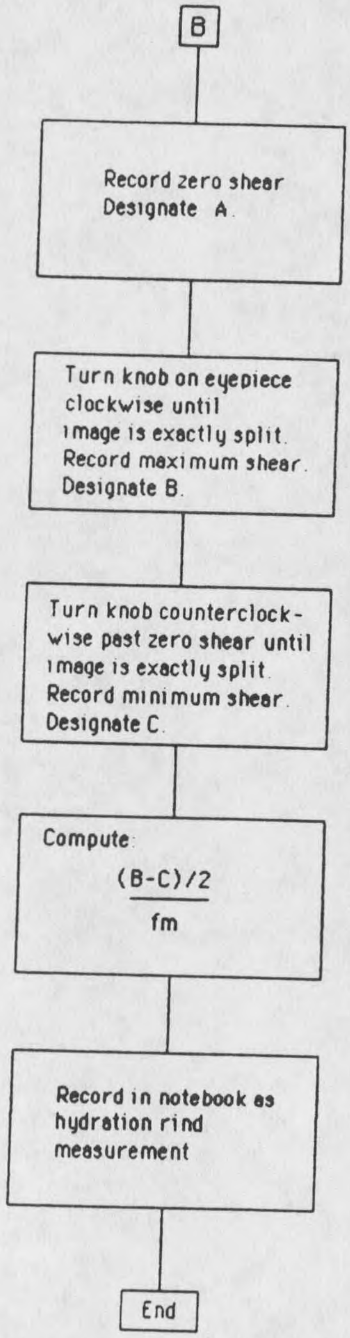


Figure 25 - (continued).



APPENDIX C

CHEMICAL ANALYSES OF OBSIDIAN GRAINS

Table 7. Chemical analyses of detrital beach pebbles from the Breeze Point area. Shoreline designations are in parenthesis. All values are expressed as % / weight (after Michels, 1988).

Specimen No.		Na_2O	K_2O	$Fe_2O_3^T$	CaO	MgO	Source
456-L87-13C-1	(S1)	3.50	5.23	1.41	0.46	0.08	Aster Creek flow
456-L87-15B-1	(S4)	3.35	5.23	1.39	0.40	0.08	Aster Creek flow
456-L87-17C-1	(S5)	3.24	5.37	1.28	0.41	0.08	Aster Creek flow
456-L87-15B-2	(S4)	3.40	5.24	1.49	0.43	0.08	Aster Creek flow
456-L87-15D	(S4)	3.34	5.59	1.38	1.05	0.12	Unident. type A
456-L87-15E	(S4)	3.18	5.54	1.39	0.47	0.08	Unident. type B
456-L87-15A-3	(S4)	3.46	5.21	1.44	0.40	0.07	Aster Creek flow
456-L87-16A	(S3)	3.45	5.33	1.47	0.41	0.07	Aster Creek flow
456-L87-17B-1	(S5)	3.59	5.16	1.48	0.48	0.09	Aster Creek flow
456-L87-16B	(S3)	3.53	5.22	1.37	0.43	0.08	Aster Creek flow
456-L87-13E	(S1)	3.38	5.44	1.38	0.44	0.08	Aster Creek flow
456-L87-17C-2	(S5)	3.35	5.51	1.44	0.49	0.09	Aster Creek flow
456-L87-17B-2	(S5)	3.54	5.31	1.47	0.44	0.08	Aster Creek flow
456-L87-13C-2	(S1)	3.46	5.24	1.46	0.47	0.09	Aster Creek flow
456-L87-17B-3	(S5)	3.52	5.56	1.54	1.02	0.14	Unident. type A
456-L87-14B	(S2)	3.52	5.26	1.48	0.50	0.12	Aster Creek flow
456-L87-15A-1	(S4)	3.43	5.26	1.42	0.46	0.09	Aster Creek flow
456-L87-15A-2	(S4)	3.45	5.34	1.46	0.55	0.13	Aster Creek flow
456-L87-17A-1	(S5)	3.21	5.50	1.43	0.49	0.10	Unident. type B
456-L87-14D-2	(S2)	3.38	5.26	1.41	0.46	0.09	Aster Creek flow
456-L87-15C	(S4)	3.44	5.19	1.41	0.50	0.10	Aster Creek flow
456-L87-14A	(S2)	3.39	5.20	1.41	0.48	0.11	Aster Creek flow
456-L87-14D-1	(S2)	3.43	5.22	1.38	0.56	0.12	Aster Creek flow
456-L87-17B-4	(S5)	3.38	5.40	1.48	0.50	0.10	Aster Creek flow

Table 8. Chemical analyses of Obsidian Cliff obsidian. All values are expressed as % / weight (After Friedman and Long, 1976; Michels, 1985).

SiO_2	Al_2O_3	Na_2O	K_2O	$Fe_2O_3^T$	CaO	MgO	TiO_2
77.1	12.14	3.92	5.07	1.27	0.38	0.04	0.09
76.7	12.50	3.30	4.60	0.43	0.48	0.09	0.07

APPENDIX D

RIND MEASUREMENT DATA FOR INDIVIDUAL SAMPLING LEVELS FOR TERRACE T0

Table 9. Rind measurement data for individual sampling levels for Terrace TO

0-20 cm	20-40	40-60	60-80	80-100
1.86	4.00	2.34	4.23	2.91
2.51	4.00	3.71	4.57	5.66
2.63	4.11	4.06	4.69	6.34
2.91	4.23	4.80	4.97	6.57
3.14	4.63	4.91	5.09	7.94
3.20	4.69	5.60	5.26	8.69
3.54	4.74	5.66	5.20	10.23
3.66	5.03	5.66	5.49	10.40
3.77	5.26	5.71	5.89	12.23
4.51	5.94	5.89	6.06	12.57
4.74	5.94	5.94	6.06	12.80
4.91	6.17	6.06	6.74	13.89
5.20	6.23	6.17	6.86	14.40
5.43	6.29	6.51	6.97	14.51
5.83	6.86	6.68	7.14	14.97
6.23	7.26	6.74	7.43	15.43
6.34	7.26	6.91	7.77	20.23
7.54	7.71	7.20	8.57	20.86
7.66	8.06	7.37	8.91	20.97
8.06	8.57	7.43	9.03	21.20
8.06	9.43	7.49	9.44	21.83
8.34	9.49	8.06	11.66	22.11
8.57	10.74	8.17	12.00	23.03
9.03	11.14	9.14	12.34	23.83
9.14	11.49	9.14	12.91	26.06
9.31	12.34	9.77	13.37	26.63
12.46	14.51	11.26	14.17	28.74
12.91	17.60	11.31	14.97	
13.26	19.54	12.06	15.43	
13.49	22.86	13.09	16.11	
15.60	23.26	14.69	18.00	
22.80		15.14	23.77	
		15.66	23.77	
		19.43		
		21.26		

APPENDIX E

RIND MEASUREMENT DATA FOR TERRACES T1 THROUGH T4 AND THE MODERN STREAM

Table 10. Hydration rind measurements for Terraces T1 through T4 and the modern stream.

T1	T2	T3	T4	Modern
0	0	0	4.23	0
0	0	0	4.40	0
2.40	3.26	4.00	4.46	4.46
3.43	3.49	4.34	4.51	5.94
4.11	4.23	4.63	4.74	6.51
4.34	4.40	4.69	4.80	6.51
4.34	4.57	4.91	4.97	6.69
4.80	4.57	5.49	5.03	6.74
4.97	4.74	5.60	5.14	6.97
5.14	4.80	5.71	5.20	7.31
5.26	4.80	5.83	5.20	7.54
5.94	5.09	5.89	5.26	8.06
6.00	5.20	6.00	5.31	8.11
6.17	5.20	6.06	5.31	8.40
6.17	5.31	6.23	5.31	8.46
6.23	5.43	6.29	5.37	8.46
6.23	5.49	6.40	5.54	9.03
6.29	5.83	6.46	5.60	9.09
6.57	5.94	6.51	5.71	9.20
6.74	6.06	6.63	5.71	9.26
7.26	6.17	6.74	5.83	9.31
7.37	6.23	6.80	6.06	9.89
7.43	6.23	6.80	6.17	10.00
7.54	6.29	7.14	6.40	10.29
7.60	6.34	7.14	6.63	10.40
7.83	6.34	7.20	6.80	11.26
7.89	6.40	7.20	6.91	11.37
7.94	6.57	7.54	7.09	11.43
8.06	6.80	7.66	7.09	11.66
8.11	6.80	7.71	7.20	11.71
8.40	6.86	7.71	7.26	11.77
8.69	6.86	8.11	7.43	11.89
8.97	6.97	8.34	7.43	12.23
8.97	7.09	8.57	7.54	12.51
9.03	7.49	9.03	7.60	12.63
10.29	7.77	9.14	7.66	12.69
10.63	7.83	9.31	7.66	12.74
10.86	8.06	9.54	7.71	13.09
11.09	8.11	10.06	7.77	13.31
11.09	8.17	10.06	7.83	13.71
11.09	8.23	10.86	7.89	14.63
11.31	8.69	11.09	8.11	15.20
11.54	9.09	11.26	8.46	15.43
11.54	10.00	11.37	8.91	15.49
11.66	10.46	11.94	9.14	15.60
11.89	10.69	12.34	9.26	16.23
12.17	11.20	12.51	10.06	16.29

Table 10. (continued).

T1	T2	T3	T4	modern
12.23	11.83	12.69	10.40	18.06
12.29	12.11	13.14	10.51	18.51
12.57	12.17	13.26	10.86	18.63
12.57	12.29	13.54	11.09	19.60
13.37	12.63	15.31	11.09	20.57
13.54	13.43	15.66	12.00	20.74
13.94	13.54	16.00	12.06	20.86
13.94	13.71	16.74	12.23	21.83
14.29	14.00	17.31	12.74	21.89
14.29	14.34	18.29	12.80	22.74
14.34	14.46	18.34	13.09	23.14
14.57	14.86	18.74	13.09	25.83
15.09	14.91	20.74	13.43	43.26
15.14	14.97	21.49	15.49	
15.54	15.54	35.43	15.54	
16.06	15.89		15.89	
16.74	17.03		15.94	
16.80	17.49		16.17	
16.86	17.60		16.80	
17.03	17.66		16.80	
17.37	18.23		17.49	
17.49	18.97		17.77	
18.00	19.37		18.11	
18.00	19.43		18.34	
18.29	19.89		18.40	
18.51	21.31		18.51	
18.69	21.49		18.69	
19.09	23.49		19.03	
19.94	23.71		19.14	
20.91	31.31		19.20	
22.74	38.58		19.54	
23.14			20.11	
24.46			20.29	
			21.09	
			21.83	
			22.63	
			23.43	
			23.49	
			24.06	
			25.37	
			25.43	
			25.71	
			26.69	
			36.63	

APPENDIX F

RIND MEASUREMENT DATA FOR SHORELINES

Table 11. Rind measurement data for Shorelines S1 through S5 at Yellowstone Lake.

S1	S2	S3	S4	S5
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	0
0	0	0	0	3.09
0	0	3.26	0	4.00
0	0	3.31	0	4.06
0	0	3.54	0	4.17
0	3.77	3.89	0	4.23
0	3.89	4.29	2.86	4.30
0	4.00	4.34	3.20	4.63
0	4.34	4.74	3.31	4.69
0	4.34	4.80	3.54	4.69
0	4.40	4.86	3.71	4.69
0	4.46	4.91	3.89	4.80
0	4.51	5.09	3.89	4.86
0	4.57	5.26	4.11	4.86
0	4.57	5.31	4.23	4.86
0	4.69	5.60	4.29	5.31
0	4.69	5.83	4.34	5.37
0	4.80	6.29	4.51	6.00
0	4.80	6.29	4.69	6.00
0	4.80	6.40	4.69	6.86
0	4.91	6.51	4.69	6.91
0	4.91	6.57	4.74	7.66
0	4.91	6.74	4.80	11.31
2.74	4.97	6.86	4.91	11.43
3.31	5.03	6.91	4.91	11.54
3.43	5.09	6.97	4.91	11.89
3.71	5.20	7.09	4.97	12.97
3.83	5.31	7.20	5.03	14.40
3.89	5.31	7.43	5.09	16.23
3.94	5.37	7.60	5.09	16.63
3.94	5.37	7.66	5.49	16.80
4.06	5.37	7.89	5.49	19.31
4.06	5.43	7.89	5.49	19.60
4.17	5.49	8.69	5.49	19.94

Table 11. (continued).

S1	S2	S3	S4	S5
4.17	5.49	8.80	5.60	19.94
4.23	5.49	9.20	5.89	19.94
4.29	5.49	9.94	6.34	20.23
4.34	5.54	10.11	6.57	20.46
4.34	5.60	11.37	6.63	20.69
4.46	5.71	11.43	6.97	20.80
4.46	5.71	11.89	7.03	21.14
4.63	5.83	12.57	7.09	21.31
4.69	5.83	12.97	7.54	21.54
4.91	5.94	13.49	7.77	21.77
4.91	5.94	13.60	8.17	21.89
4.91	5.94	14.17	8.97	22.34
4.97	6.00	14.80	9.49	22.51
5.09	6.11	14.97	10.06	22.63
5.14	6.17	15.54	10.17	22.63
5.14	6.17	15.60	10.80	24.00
5.20	6.29	15.60	11.03	25.14
5.31	6.34	15.66	11.31	25.43
5.31	6.46	15.71	11.31	
5.37	6.46	15.71	12.23	
5.49	6.86	15.77	13.43	
5.83	7.03	15.89	13.71	
6.06	7.09	16.11	14.86	
6.06	7.14	16.11	15.26	
6.46	7.26	16.46	15.37	
6.69	7.49	16.51	16.00	
6.69	8.51	16.57	16.06	
7.20	9.14	16.57	16.40	
7.37	9.31	16.57	16.46	
7.54	9.31	16.80	16.57	
7.89	10.40	17.14	17.03	
7.89	10.40	17.43	17.54	
7.89	10.97	17.66	17.54	
8.23	11.09	17.77	23.26	
9.60	11.26	18.57		
9.66	11.37	18.86		
9.89	11.49	19.20		
9.94	11.66	19.31		
10.17	11.66	19.60		
10.29	13.03	19.66		
10.40	13.03	20.06		
11.14	13.14	20.74		
11.37	13.49	21.14		
11.89	14.17	21.60		
12.34	15.43	23.77		
12.46	16.46	26.51		
12.80	16.46	29.60		
13.94	16.80	30.34		

Table 11. (continued).

S1	S2	S3	S4	S5
14.00	16.91			
14.51	17.26			
21.71	17.66			
26.74	21.43			
	23.20			
	24.11			
	30.06			