

# PROCEEDINGS OF THE TWENTY-EIGHTH ANNUAL KECK RESEARCH SYMPOSIUM IN GEOLOGY

April 2015  
Union College, Schenectady, NY

Dr. Robert J. Varga, Editor  
Director, Keck Geology Consortium  
Pomona College

Dr. Holli Frey  
Symposium Convener  
Union College

Carol Morgan  
Keck Geology Consortium Administrative Assistant

Christina Kelly  
Symposium Proceedings Layout & Design  
Office of Communication & Marketing  
Scripps College

*Keck Geology Consortium  
Geology Department, Pomona College  
185 E. 6<sup>th</sup> St., Claremont, CA 91711  
(909) 607-0651, keckgeology@pomona.edu, keckgeology.org*

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Robert J. Varga  
Editor and Keck Director  
Pomona College

Keck Geology Consortium  
Pomona College  
185 E 6<sup>th</sup> St., Claremont, CA  
91711

Christina Kelly  
Proceedings Layout & Design  
Scripps College

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EMILY DIENER, Macalester College  
Research Advisor: Kelly MacGregor

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# MINERALOGICAL AND CHEMICAL COMPOSITION OF SEDIMENT FROM LOWER GRINNELL LAKE, GLACIER NATIONAL PARK, MONTANA

ASHLEIGH COVARRUBIAS, California State University, San Bernardino

Research Advisor: Joan Fryxell

## INTRODUCTION

Understanding connections between climate change and paleoenvironmental variability is fundamental for predicting the impacts of anthropogenic climate change on landscapes. The high sensitivity of alpine glaciers to climate variation (temperature and precipitation) means that glacier size can be a good proxy for reconstructing past climate variability (e.g. Hall and Fagre, 2003). However, terrestrial records of glacier position such as moraines are not always present or preserved in alpine landscapes (Carrara, 1987; Rodbell et al., 2008). The production of sediment in subglacial and supraglacial environments, and the subsequent transport of this sediment into proglacial sinks such as lakes, can provide more continuous records of the proximity of ice (Leonard, 1986; Leonard and Reasoner, 1999; Rodbell et al., 2008). For example, a larger glacier footprint can be more effective in producing subglacial sediment than a small glacier (e.g. Hallet et al., 1996), as well as decreasing transport distance to a downvalley lake. In addition, an increase in temperature may increase glacial sediment flux by increasing glacier sliding, subglacial abrasion, and melt water flux to transport the sediment (Longwell and Flint, 1962). In valleys where different rock types are present, the mineralogy and/or geochemical fingerprint of sediment deposited in proglacial lakes may reveal which part of the landscape was being actively eroded over time.

Glacier National Park, Montana, is an ideal place to study the relationship between glacier size, climate, and sediment transport. In Glacier National Park as a whole, the number of valley glaciers has decreased from 150 in 1850 (end of Little Ice Age) to ~25

glaciers today (Dyson, 1941; <http://pubs.usgs.gov/of/1993/0510/report.pdf>, p.18; Hall and Fagre, 2003). Since the mid-19th century, Grinnell Glacier had covered up to about 500 acres, but by 1984 only spanned 225 acres of its cirque basin (Carrara, 1993). Presently, Grinnell Glacier has retreated into its cirque basin of approximately <220 acres, and is eroding the stromatolitic Siyeh Limestone of the Helena Formation (Key et al., 1994). Previous work has suggested that the presence of dolomite in a sediment core from Swiftcurrent Lake is evidence for a more extensive Grinnell Glacier (MacGregor et al., 2011; Schachtman et al., in press). However, limited work on the mineralogical and geochemical characteristics of the rock units has made it difficult to constrain the source of lake sediments with any confidence. This study aims to identify the mineralogical and geochemical signatures of a 1,200-year long sediment core from lower Grinnell Lake in an effort to determine if geomorphic sources of eroded rock vary during that time.

## STUDY SITE

Many Glacier is within the Lewis Range and is located in the northeastern portion of Glacier National Park, roughly 21 km southwest of Babb, Montana. Lower Grinnell Lake lies about five km up Grinnell Glacier valley, which is located less than three km east of the Continental Divide and the Garden Wall headwall (MacGregor et al., 2011; Griffith et al., 2011). Upper Grinnell Lake is the most proximal proglacial lake to both retreating Salamander and Grinnell Glaciers within the valley, and is currently in contact with Grinnell Glacier in its cirque basin. Beyond the cirque basin, water flows from upper Grinnell Lake

downvalley through lower Grinnell Lake (study site; Fig. 1 of project summary), Lake Josephine, and Swiftcurrent Lake, before leaving the valley.

The Grinnell Glacier valley basin is composed of thick sequences of the mid Proterozoic Belt-Purcell Supergroup, which consists mostly of argillaceous, quartzitic, and carbonaceous strata (Ross, 1951). Lower Grinnell Lake is the catchment basin directly beneath Grinnell Glacier with about a 2 km difference in distance and a change in elevation of around 460 meters (MacGregor et al., 2011). The lake presently receives minerals and glacial meltwater directly from a tributary of the glacial cirque basin (Osborn et al., 1995). With lower Grinnell Lake being directly downstream of the eroding glacier, the erosion of the stromatolitic Siyeh Limestone of the Helena formation is the only source of bedrock that contains dolomite within the valley (Whipple, 1992). Lower Grinnell Lake overlies the mid Proterozoic Belt Supergroup that is predominantly composed of carbonatic formations such as the Appekunny Argillite, Grinnell Argillite, and Empire Formation, which includes siltstones, shales, and sandstones (MacGregor et al., 2011). The valley walls surrounding the glacier are comprised of the Middle Belt-Purcell Helena Formation that consists of dolomitic and siliciclastic facies, the Siyeh Limestone, along with the igneous Purcell Sill (Ross, 1951; Gonzalez-Alvarez and Kerrich, 2011; MacGregor et al., 2011).

Previous paleoenvironmental investigations of the mid Proterozoic Belt-Purcell Supergroup formations have shown that the rocks are dominantly siliciclastic and carbonate-rich rocks that were deposited between 1470 and 1400 million years ago and reaches up to about 750 meters in thickness (Gonzalez-Alvarez and Kerrich, 2011). Lower Belt rocks were deposited in deep waters with marginal shoreline facies that evolved into the shallow and tidal areas that is known as the Ravalli Group, which consists of the Empire, Grinnell, and Appekunny Formations. The Middle Belt environment developed into a shallow carbonate platform that became exposed and resulted in terrestrial deposits that is known as the Mid-Purcell Group that consists of the Helena Formation, Siyeh Limestone, and Purcell Lava, as well as the Snowslip and Shepard Formations within the Missoula Group

(Gonzalez-Alvarez and Kerrich, 2011; Griffith et al., 2011; MacGregor et al., 2011).



Figure 1. In 2010 the selected cores from lower Grinnell Lake LGRIN10-1B-1P-1 (upper right portion) and LGRIN10-1C-2B-1 (lower left portion) were photographed at LacCore. Above shows the adjacent overlapping cores with a total of 158 cm that was collected for XRD and XRF analysis. The core photographs are courtesy of LacCore, University of Minnesota (Griffith et al., 2011).



## METHODS

Correlative, adjacent overlapping cores LGRIN10-1B-1P-1 (0.0-0.82 m sediment depth) and LGRIN10-1C-2B-1 (0.0-4.68 m sediment depth) (Fig. 1) were collected in the summer of 2010 (Griffith et al., 2011). Each core was taken back to LacCore at the University of Minnesota where the cores were logged for bulk density and magnetic susceptibility. Each core was split into archival and working halves, digitally photographed, and described using initial core descriptions (ICDs) and smear slides. The samples collected from core LGRIN10-1B-1P-1 included 1 cm intervals from 15 cm to 75 cm, for a total of 60 cm. The samples collected from core LGRIN10-1C-2B-1 included 1 cm intervals from 38 cm to 136 cm, for a composite total of 158 cm. Subsampling of 4 cc at 1 cm intervals on the cores took place at the LacCore lab, and then the cores were wrapped and refrigerated at 4° C for future work (Myrbo, 2004a). Geochemical analyses were conducted at Pomona College and California State University, San Bernardino.

The procedures for determining mineralogical and chemical composition of lower Grinnell Lake include X-ray diffraction (XRD) that determines minerals present in the sediment by crystal structure and X-ray Fluorescence (XRF), which analyzes the major and minor elements within the sediment. The process for both XRD and XRF consisted of freeze drying (Myrbo, 2004b) all 127 samples at the University of Minnesota and crushing each sample in a shatter box at CSUSB and Pomona College.

The process for XRF was conducted at Pomona College using Whole Rock Analysis: XRF X-ray Fluorescence Panalytical Axios: 2010. Powdered samples were weighed with a 2:1 ratio of core sediment used to the ratio of flux used (anhydrous,

vitreous mixture of lithium tetraborate and lithium metaborate powder), placed in graphite crucibles, and fused in a furnace at 1000° C to completely melt the powdered mixture. The samples were taken out to cool down and form into glass beads that were then crushed again and placed in a furnace at 1000° C for a second fusion. After this process, the glass beads were then polished with meshes of 100, 320, 600, 1200, and 3000 grits, placed in a sonic bath of ethanol for about two minutes and engraved with the proper identification indicator label. Details of analytical and preparation processes are available at: <<https://sites.google.com/site/xrpfomona/home/sample-preparation>>. After the preparation stage for each bead, they were then placed and analyzed in the XRF machine for elemental composition along with loss on ignition (LOI) to measure % total carbon.

Analysis for XRD was conducted at CSUSB using the Panalytical X'Pert Pro X-ray Powder Diffractometer through X'Pert Data Collector and data was analyzed through the X'Pert High Score interpretation program. Powder samples were crushed in a small agate crucible, placed on a microscope slide, and put into the XRD machine for up to 4 to 5 hours on a wide scan of 10 to 90 degrees for crystal structure analysis. The minerals for XRD that are concentrated on include calcite, quartz, dolomite, and phyllosilicate minerals such as micas and illite. More information for XRD is available at: <<http://nsci.csusb.edu/physicslab/facilities.html>>.

We used the age-depth model developed by Griffith et al. (2011) to show the lake record of linear interpolation of accelerator mass spectrometry (AMS) radiocarbon ( $^{14}\text{C}$ ) dates from approximately 1200 cal yr B.P. to present (Griffith et al., 2011). The age-depth model used allows us to identify important environmental, geological, and anthropogenic events that might have happened within this time frame.

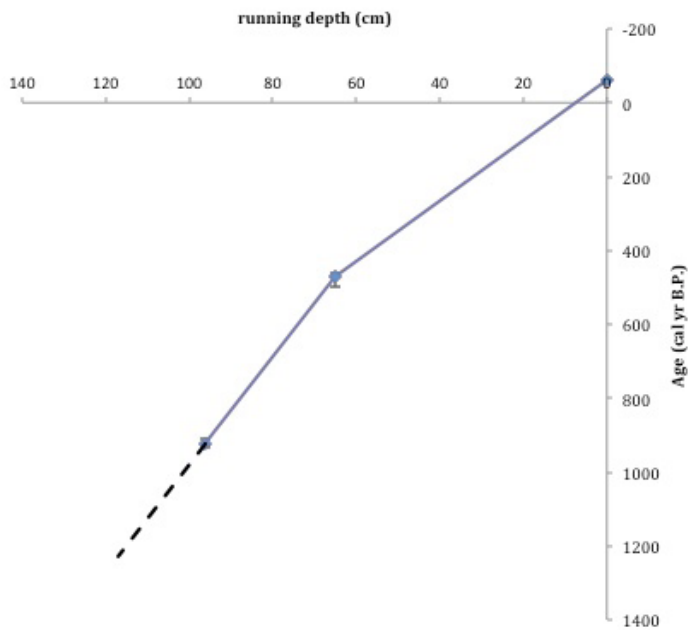


Figure 2. An age-depth model constructed by Jonathan Griffith in 2011 for Grinnell Lake shows linear interpolation of radiocarbon dates from AMS that were calibrated into calendar years (cal yr B.P.). Griffith notes from the diagram that the sedimentation rate from approximately 0-470 cal yr B.P. was 0.138 cm/yr, and the sedimentation rate from approximately 470-1200 cal yr B.P. was 0.069 cm/yr (Griffith et al., 2011).

## RESULTS

The dolomitic units in the Park consist of quartz, calcite, micas, and clay minerals, where dolomitic crystals of fine and coarse variation are mixed with argillitic sediment (e.g. Gonzalez-Alvarez and Kerrich, 2011). The smear slide analysis indicates the sediment is primarily quartz, dolomite, and calcite, with negligible amounts of trace minerals. Mineralogy was fairly consistent throughout the continuous core.

## XRF

XRF results are shown as both a function of composite core depth and age in years before present (ybp) (Fig. 2; Fig. 3). Geochemical analyses include %SiO<sub>2</sub>, %Al<sub>2</sub>O<sub>3</sub>, and rubidium (Rb), barium (Ba), zirconium (Zr), zinc (Zn), and sulfur (S) in ppm, along with % carbon loss-on-ignition (LOI).

SiO<sub>2</sub> is fairly stable throughout the period of record, with an average value of ~60% (Fig. 3a). Al<sub>2</sub>O<sub>3</sub>

appears to increase from ~1.1-1.6% from 1100-500 ybp, then decreases with a slight jump about 100 ybp (Fig. 3b). Ba increases between 1100 and present, with a range of high values around 600 ybp (Fig. 3c). Zr increases from 160 to 225 ppm between 1100 and 500 ybp, and then drops to ~160 ppm until present (Fig. 3d). S is highly variable between 560 and 620 ybp, with a trend of increasing average S from 1100 ybp and present (Fig. 3e). Percent total carbon decreases from 14-4% between 1100 and ~650 ybp, increases and then decreases abruptly around 600 ybp, and increases toward present (Fig. 3f).

There are some periods of time in which several geochemical signals vary simultaneously. The end of the Little Ice Age (LIA) around 175-ybp is coincident with a peak in Al<sub>2</sub>O<sub>3</sub> and Ba, and a decrease in S. Between ~425 and 175 ybp (gray box) most measurements show low variability and are generally stable. Between ~650 and 425 ybp, all variables except % total carbon increase to their highest values, with peaks at 600 ybp. Between 1100 and ~650 ybp, all variables (except % carbon) increase very slightly, but are generally stable.

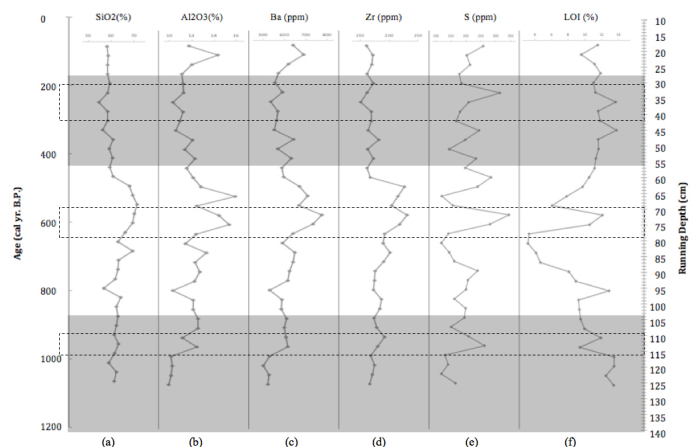


Figure 3. X-ray fluorescence (XRF) results showing major and minor trace elements within both lake cores of lower Grinnell Lake LGRIN10-1B-1P-1 and LGRIN10-1C-2B-1 in correlation of relative depth from Griffith (2011) age-depth model. (a) percent silicon dioxide (b) percent aluminum oxide (c) parts per million barium (d) parts per million zirconium (e) parts per million sulfur (f) percent loss on ignition. Gray areas highlight approximate glacial periods and black dashed lines represent interesting patterns through core sediment.



## X-Ray Diffraction (XRD)

XRD analysis identified quartz ( $\text{SiO}_2$ ), dolomite ( $\text{CaMg}(\text{CO}_3)_2$ ), and calcite ( $\text{CaCO}_3$ ) in core sediments. The minor minerals identified as a whole included ankerite ( $\text{Ca}(\text{Fe}, \text{Mg}, \text{Mn})(\text{CO}_3)_2$ ), graphite (C), spinel ( $\text{MgAl}_2\text{O}_4$ ), and magnetite ( $\text{Fe}^{2+}, \text{Fe}^{3+}_2\text{O}_4$ ). The clay/ phyllosilicate minerals included illite ( $(\text{Al}, \text{Mg}, \text{Fe})_2(\text{Si}, \text{Al})_4\text{O}_{10}[(\text{OH})_2, (\text{H}_2\text{O})]$ ), montmorillonite ( $(\text{Na}, \text{Ca})_{0.33}(\text{Al}, \text{Mg})_2(\text{Si}_4\text{O}_{10})(\text{OH})_2 \cdot n\text{H}_2\text{O}$ ), clinochlore ( $(\text{Mg}_5\text{Al})(\text{AlSi}_3\text{O}_{10})(\text{OH})_8$ ), and micas such as biotite ( $\text{K}(\text{Mg}, \text{Fe})_3\text{AlSi}_3\text{O}_{10}(\text{F}, \text{OH})_2$ ), muscovite ( $\text{KAl}_2(\text{AlSi}_3\text{O}_{10})(\text{F}, \text{OH})_2$ ), and lepidolite ( $\text{K}(\text{Li}, \text{Al})_3(\text{Al}, \text{Si}, \text{Rb})_4\text{O}_{10}(\text{F}, \text{OH})_2$ ). For the sediment samples, quartz was the major mineral found in all samples analyzed, followed by abundant amounts of dolomite, calcite, ankerite, and illite/ clay minerals. Dolomite is present in greater abundance than calcite in all samples younger than 1000 ybp, although dolomite was not present in samples at 108-109 cm. Illite and clay minerals are present in all samples. Clay minerals that consist of micas such as muscovite are present in various intervals of 27-28 cm, 57-58 cm, 63-64 cm, 68-69 cm, and 72-73 cm. The minor minerals such as magnetite ( $\text{Fe}_3\text{O}_4$ ) or hematite ( $\text{Fe}_2\text{O}_3$ ) showed up in intervals 17-18 cm, 29-30 cm and lower intervals, as well as perovskite ( $\text{CaTiO}_3$ ), which showed up in 20-21 cm, 50-51 cm, and 108-109 cm.

## DISCUSSION

Variability in the analyzed core samples indicate changes in the source rock delivering sediment to lower Grinnell Lake. Work by another student (Stephens et al., this volume) shows that Ba and Zr are found in all of the formations in the valley, and may not be good indicators of one stratigraphic unit. Because quartz and clays are major components of all rock types in the valley,  $\text{SiO}_2$  and  $\text{Al}_2\text{O}_3$  may also be of limited use in determining changes in source rock over time. However, the coincident changes in several variables, particularly the increase found ~600 ybp, suggests there are some differences in the geochemistry of rock contributions to lake sedimentation over time. Although it may not be possible to distinguish sources of sediment using these geochemical tracers without further work, it is likely that sediment flux and stratigraphic sources

vary between 1100 ybp and present. Grinnell Glacier dynamics, and proximity to the lake, likely play an important role. Sediment variability is directly related to periods of increased clastic sediment that reflect glacial advances, warming periods, and any heavy precipitation events (Rodbell et al., 2008; Griffith et al., 2011).

Glacial fluctuation periods such as the Medieval Climate Anomaly (MCA) can be seen at approximately 1215 ybp (Fig. 3)(Griffith et al., 2011). The interval of 103-125 cm within the gray area indicates a cooler period plus an interesting peak within S and percent  $\text{Al}_2\text{O}_3$  around 975 ybp. The end of the LIA is apparent near the period of approximately 915 ybp (within 1000 to 900 ybp) on figure 3. The period from approximately 895 to 800 ybp indicates a steady period but around 800 ybp,  $\text{SiO}_2$ ,  $\text{Al}_2\text{O}_3$ , Ba, S, and % carbon display distinct decreasing peaks. The period from approximately 620 to 560 ybp within the dashed rectangle focuses on a greater influx within the minerals present. An interesting peak around 600 ybp is seen distinctly within the sediment that could have been influenced by glacial meltwater, therefore increasing mineral presence. During 425 to 175 ybp, it seems to be a much colder period where not much activity from glacial melt is apparent, but could have been an increase in precipitation. At the end of the LIA at around 175 ybp, the temperature increases which relates to a sediment influx and glacial recession, which can also be seen to the present.

With glacier size diminishing and meltwater becoming more prominent, especially within summer months, it leaves plenty of room for increased glacial flour and sediment transportation downvalley through each lake (Hall and Fagre, 2003; Rodbell et al., 2008). Therefore, with Grinnell Glacier currently eroding the Siyeh Limestone of the Helena Formation and it being the only source of dolomite within the valley, we can determine that solar forcing and climate change patterns of glacial retreat are responsible for the major and minor trace elements that are deposited within the glacial proximate lakes (Leonard, 1986; Hallet et al., 1996; Rodbell et al., 2008; MacGregor et al., 2011; Griffith et al., 2011).

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