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# **DEVELOPMENT OF A LOCAL METEORIC WATER LINE FOR SOUTHEASTERN IDAHO, WESTERN WYOMING, AND SOUTH- CENTRAL MONTANA**



U.S. Geological Survey  
Scientific Investigations Report 2004-5126

**Cover: From left to right, Mount Owen, Grand Teton, and Middle Teton looking east from Table Mountain, Wyoming. Note small glacier in foreground. (Photograph© by Gordon W. Rattray, Ririe, Idaho. Used with permission.)**

# **DEVELOPMENT OF A LOCAL METEORIC WATER LINE FOR SOUTHEASTERN IDAHO, WESTERN WYOMING, AND SOUTH-CENTRAL MONTANA**

By Lyn Benjamin, Consulting Hydrologist, LeRoy L. Knobel, U.S. Geological Survey, L. Flint Hall, State of Idaho  
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**U.S. GEOLOGICAL SURVEY**

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**U.S. DEPARTMENT OF ENERGY**

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## CONVERSION FACTORS, VERTICAL DATUM, AND DELTA NOTATION

Multiply	By	To Obtain
centimeter (cm)	0.3937	inch
kilometer (km)	0.6214	mile
liter (L)	0.2642	gallon
meter (m)	3.281	foot
milliliter (mL)	0.002113	pint
square kilometer (km <sup>2</sup> )	0.3861	square mile

For temperature, degrees Celsius (°C) may be converted to degrees Fahrenheit (°F) by using the formula  
 $^{\circ}\text{F} = (1.8) (^{\circ}\text{C}) + 32$ .

Sea Level: in this report, “sea level” refers to the National Geodetic Vertical Datum of 1929, a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called “Sea Level Datum of 1929.”

Abbreviated units used in report:  $\mu\text{S}/\text{cm}$  (microsiemens per centimeter at 25°C),  $\text{mg}/\text{L}$  (milligrams per liter),  $\mu\text{g}/\text{L}$  (micrograms per liter).

### Delta Notation for Reporting Stable Isotope Data

The absolute measurement of isotopic ratios is a difficult analytical task and, as a result, relative isotopic ratios are measured as a matter of convention (Toran, 1982). For example,  $^{18}\text{O}/^{16}\text{O}$  of a sample is compared with  $^{18}\text{O}/^{16}\text{O}$  of a standard:

$$\delta^{18}\text{O} = (R_{\text{sample}}/R_{\text{standard}} - 1) \times 1,000,$$

where

$$R_{\text{sample}} = ^{18}\text{O}/^{16}\text{O} \text{ in the sample,}$$

$$R_{\text{standard}} = ^{18}\text{O}/^{16}\text{O} \text{ in the standard, and}$$

$\delta^{18}\text{O}$  = relative difference in concentration, in units of parts per thousand (permil).

Delta  $^{18}\text{O}$  ( $\delta^{18}\text{O}$ ) is referred to as delta notation and is the value reported by isotopic laboratories for stable isotope analysis. Delta  $^2\text{H}$  ( $\delta^2\text{H}$ ) can be derived by analogy to  $\delta^{18}\text{O}$  where the ratio  $^2\text{H}/\text{H}$  replaces  $^{18}\text{O}/^{16}\text{O}$  in  $R_{\text{sample}}$  and  $R_{\text{standard}}$ . The standard used for determining  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  in water originally was standard mean ocean water (SMOW) as defined by Craig (1961). The standard used in this report is Vienna standard mean ocean water (VSMOW). If  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  samples contain more of the heavier isotopes ( $^{18}\text{O}$  or  $^2\text{H}$ ) than the reference material, the samples have positive permil values and are referred to as heavier than the reference material or as being enriched in the heavier isotope. Conversely, if the samples contain more of the lighter isotopes ( $^{16}\text{O}$  or  $\text{H}$ ) than the reference material, the samples have negative permil values and are referred to as lighter than the reference material or as being depleted in the heavier isotope. For example, a  $\delta^{18}\text{O}$  value of -18.15 can be referred to as lighter than VSMOW or depleted in  $^{18}\text{O}$  relative to VSMOW. Once the reference material has been specified, it is assumed by convention that all values are reported relative to it unless otherwise indicated. Because VSMOW reflects the average isotopic composition of the ocean, and because of the nature of isotope fractionation processes,  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  values of precipitation are always negative. The same terminology for discussions of  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  relative to VSMOW can be applied to different samples of precipitation that have different values. For example, if two samples of precipitation have  $\delta^2\text{H}$  values of -144.9 and -150.5, then the sample with the value of -150.5 can be referred to as lighter than the sample with the value of -144.9. In a similar fashion, the sample with the value of -150.5 is depleted in the heavier isotope relative to the other sample.

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## Abstract

Linear-regression analysis was applied to stable hydrogen (H) and oxygen (O) isotope data in 72 snow-core and precipitation samples collected during 1999-2001 to determine the Local Meteoric Water Line (LMWL) for southeastern Idaho, western Wyoming, and south-central Montana.

On the basis of (1) residuals from the regression model, (2) comparison of study-area deuterium-excess (*d*-excess) values with the global range of *d*-excess values, and (3) outlier analysis by means of Chauvenet's Criterion, values of four samples were excluded from final regression analysis of the dataset. Regression results for the 68 remaining samples yielded a LMWL defined by the equation  $\delta^2\text{H} = 7.95 \delta^{18}\text{O} + 8.09$  ( $r^2 = 0.98$ ).

This equation will be useful as a reference point for future studies in this area that use stable isotopes of H and O to determine sources of ground-water recharge, to determine water-mineral exchange, to evaluate surface-water and ground-water interaction, and to analyze many other geochemical and hydrologic problems.

## INTRODUCTION

As a result of kinetic and equilibrium processes during evaporation from the ocean and subsequent condensation, the ratios of the stable isotopes of hydrogen (H and  $^2\text{H}$  or deuterium) and oxygen ( $^{16}\text{O}$  and  $^{18}\text{O}$ ) in water within a particular airmass vary with temperature during condensation and with relative humidity during evaporation (Clark and Fritz, 1997). The stable isotope ratios of water vapor in an airmass reflect the origin of the airmass, and the ratios in the precipitation that evolves from the airmass reflect both the origin of the airmass and the conditions under which condensation occurs. As an airmass travels away from the ocean (or other source areas for water vapor) and precipitation occurs, precipitation that is enriched in the heavier isotopes leaves the airmass first. The remaining water vapor then is composed of lighter isotopes. Subsequent precipitation has an increasingly lighter stable isotope composition.

This depletion effect has been called the "continental effect" and results in lighter stable isotope ratios farther away from the ocean. Studies in California and Nevada have shown gradients in delta hydrogen-2 ( $\delta^2\text{H}$ ) composition of precipitation that range from 20 permil per 100 km close to the coast to 2 permil per 100 km farther inland (Friedman and Norton, 1970; Ingraham and Taylor, 1991; Williams and Rodoni, 1997). Furthermore, a strong linear correlation exists between mean annual isotopic composition of precipitation and mean annual surface air temperature. This relationship corresponds to a 1-permil decrease in mean annual delta oxygen-18 ( $\delta^{18}\text{O}$ ) with a 1.1 to 1.7°C decrease in mean annual temperature;  $\delta^2\text{H}$  varies with temperature in a similar manner to  $\delta^{18}\text{O}$  (Clark and Fritz, 1997, p. 64). As a result, precipitation at higher latitudes has a lighter stable isotope composition than precipitation closer to the equator. This temperature effect also is seen as a result of elevation; cooler temperatures at higher elevations result in  $\delta^{18}\text{O}$  depletion that varies between -0.15 and -0.5 permil per 100 m rise in elevation (Clark and Fritz, 1997). As a result of seasonal differences in temperature, strong seasonal variability in stable isotopic composition of precipitation occurs that is particularly pronounced in continental locations where seasonal temperature differences are extreme.

The relation between  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  in precipitation is described by the Global Meteoric Water Line (GMWL) developed by Craig (1961) and expressed by the equation:

$$\delta^2\text{H} = 8 \delta^{18}\text{O} + 10 \text{ permil.} \quad (1)$$

This relation was developed as an average of many local water lines that differ from the GMWL as a result of climatic and geographic factors. Differential fractionation of  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  occurs as a function of humidity during primary evaporation of water vapor from the ocean and as a function of temperature during secondary evaporation as rain falls from a cloud. These two factors affect the slope and intercept of the Local Meteoric Water Line (LMWL) and produce a different LMWL at different locations. For example, in arid climates such as Bahrain, where the secondary evaporation effect is especially pronounced, the equation for the LMWL (Clark and Fritz, 1997, p. 51) is:

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$$\delta^2\text{H} = 6.3 \delta^{18}\text{O} + 11.6 \text{ permil.} \quad (2)$$

and oxygen isotope analysis; and (3) use the isotopic data to establish the LMWL.

Greater isotopic fractionation of  $^{18}\text{O}$  than  $^2\text{H}$  with evaporation during rainfall or sublimation during snowfall results in disproportional enrichment of  $^{18}\text{O}$  relative to  $^2\text{H}$  and a lower slope for the LMWL. Defining the LMWL for precipitation is an important part of ground-water investigations that compare isotopic ratios in ground water or surface water with precipitation at specific locations.

For example, to define the processes that affect ground-water chemistry at and downstream from the Idaho National Engineering and Environmental Laboratory (INEEL, fig. 1), it is necessary to understand the natural geochemistry of water entering the eastern Snake River Plain aquifer (ESRPA). Ground water at the INEEL is derived from rapidly infiltrating surface water from the Big Lost River, from ground water derived from the Little Lost River and Birch Creek tributary basins (fig. 1), and from ground water moving into the area from the northeast. Recent and ongoing studies either have examined or currently are examining several aspects of ground-water geochemistry such as rock chemistry, water chemistry, mineralogy, stable isotopes, and naturally occurring radioactive isotopes (Rightmire and Lewis, 1987; Wood and Low, 1988; Mann and Low, 1994; Knobel and others, 1997; Cecil and others, 1999; Benjamin, 2000; Cecil, 2000; Cecil and others, 2000; Coplen and Kendall, 2000; Busenberg and others, 2001; Carkeet and others, 2001; Chapelle and others, 2002; Swanson and others, 2002, 2003). Although studies of the geochemistry of the ground water have utilized stable isotope data from basins tributary to the ESRPA, no stable isotope data for precipitation have been collected from basins north or northeast of the INEEL. To provide a more complete understanding of the geochemistry of water entering the ESRPA, it is necessary to know the stable isotope composition, particularly the stable isotopes of hydrogen and oxygen, of precipitation falling in the region.

With information about the stable hydrogen and oxygen isotope ratios of local precipitation, it may be possible to identify recharge areas in tributary basins and processes that occur during recharge and to evaluate surface- and ground-water interaction and many other geochemical and hydrologic problems. The establishment of a LMWL for southeastern Idaho, western Wyoming, and south-central Montana provides a baseline for comparison in future stable isotope studies for this region.

### Purpose and Scope

The purpose of this report is to describe the development of a local meteoric water line for southeastern Idaho, western Wyoming, and south-central Montana. The specific objectives of the study described in this report were to (1) identify and establish a network of precipitation collection sites in the study area that can be used for future stable isotope monitoring; (2) collect quarterly samples of precipitation for stable hydrogen

### Description of Study Area

The study area comprises approximately 14,200 km<sup>2</sup>, mostly in southeastern Idaho, but also in small parts of northwestern Wyoming and south-central Montana (fig. 1). The area includes the Lost River, Lemhi, Bitterroot, Snake River, and Teton Ranges; the Pioneer, Beaverhead, and Centennial Mountains; part of the eastern Snake River Plain; the Island Park Caldera; and the Yellowstone Plateau volcanic field. Drainage basins in the study area include: Big Wood River, Big Lost River, Little Lost River, Birch Creek, Medicine Lodge Creek, Camas Creek, Madison River, Henrys Fork of the Snake River, Teton River, and the upper Snake River (fig. 1). Water from the Big Lost River, Little Lost River, Birch Creek, and Medicine Lodge Creek drainages reaches the ESRPA mostly as ground-water underflow rather than surface-water flow. The area is characterized by north- to northwest-trending mountain ranges in the northern Basin and Range Province, the Yellowstone and Island Park Calderas, and the north/south-trending Teton Range. The highest elevation in the study area is Borah Peak in the Lost River Range (3,840 m above sea level) and the lowest elevation is at Mud Lake on the ESRP (1,455 m). All of the mountain ranges in the study area (fig. 1) contain peaks that exceed 3,050 m; however, the basins draining the northern Basin and Range Province and the Beaverhead and Centennial Mountains descend onto the sinks and playas of the ESRP, where elevations are less than 1,500 m.

### Regional Geology

The Lost River, Lemhi, and Bitterroot Ranges make up the northern part of the Basin and Range province and the western part of the study area. These long, north- or northwest-trending ranges are separated by equally long valleys and are bounded by normal faults (Maley, 1987). The ranges are composed of consolidated sedimentary strata consisting mostly of limestone, quartzite, sandstone, and shale that have been folded and faulted.

The north-central part of the study area lies east of the Beaverhead Mountains, south of the Continental Divide, and north of the ESRP. The principal topographic features are the southern part of the Beaverhead Mountains and the Centennial Mountains. The surficial rocks in this area were mapped and described as silicic volcanic rocks by McKee (1972, fig. 16-3, p. 259). Discharge from this area to the ESRPA is both ground- and surface-water flow from the Medicine Lodge and Camas Creek drainages.

The largest physiographic feature in the study area is the ESRP. The southwesterly movement of the North American Plate over a stationary hot spot beneath the Earth's crust formed the ESRP. The hot spot generated a series of volcanic fields

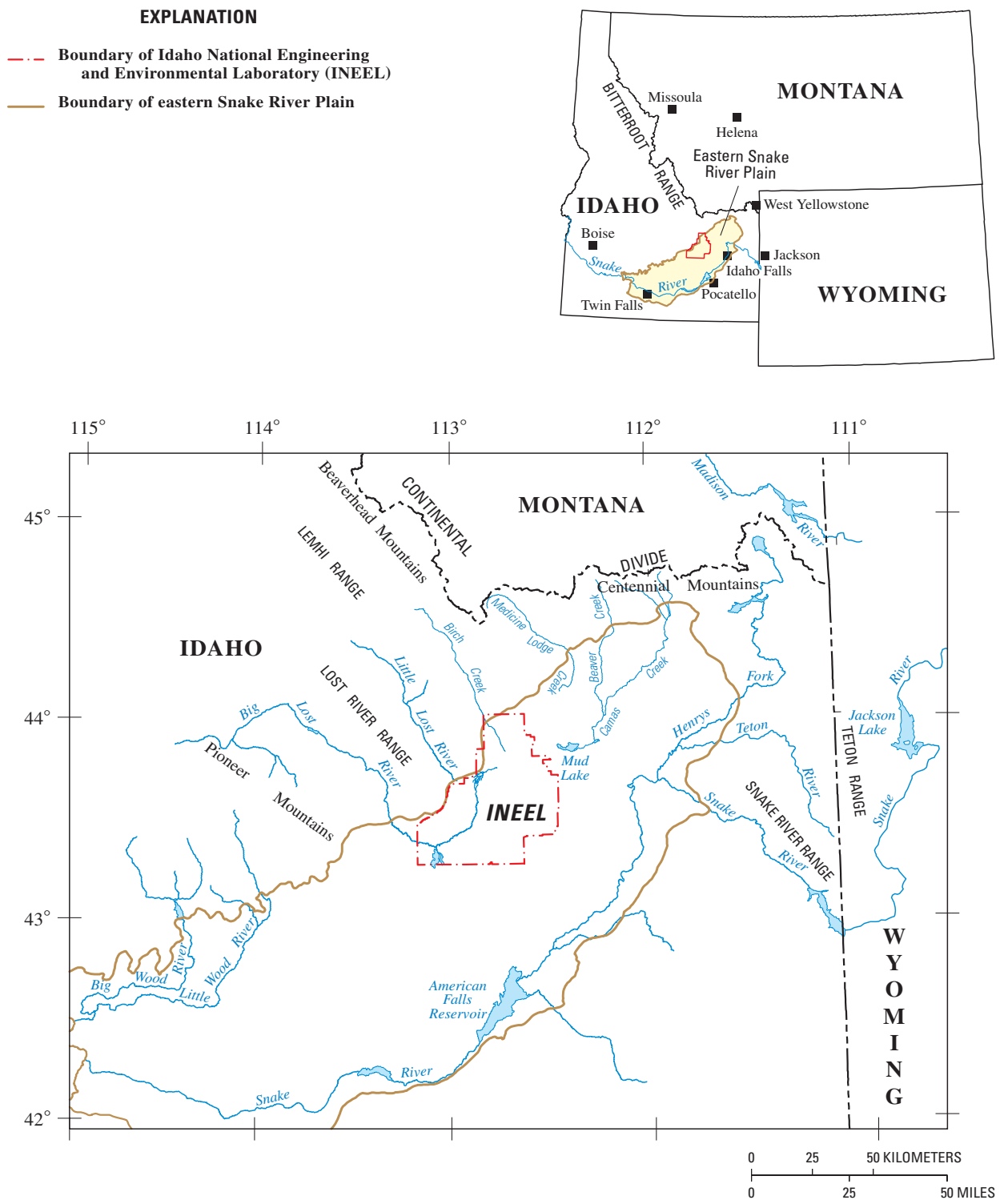


Figure 1. Location of study area, Idaho National Engineering and Environmental Laboratory, and the eastern Snake River Plain.

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that deposited silicic volcanic rocks over large areas. After the crustal material moved past the hot spot, secondary basaltic volcanism covered the silicic rocks. The basalt flows were intermittent, and lacustrine, fluvial, and eolian sediments were deposited between volcanic events. Several volcanic fields of this type have been identified by Pierce and Morgan (1992). Island Park Caldera is located at the northeastern margin of the ESRP and forms a topographic and geologic transition from the basalt of the ESRP to the Yellowstone Plateau volcanic field (Christiansen, 1982; Christiansen and Embree, 1987).

The Teton Range forms the eastern boundary of the study area and also the eastern margin of the Basin and Range Province in this area. The Teton Range is a slice of basement rock lifted along a fault that defines the eastern front of the range. If the trend of the Teton Range is projected north-south, it aligns approximately with Yellowstone National Park (YSNP) and the Madison and Gallatin Ranges of Montana to the north and with the Wasatch Range of Utah to the south (Alt and Hyndman, 1989). The upper Snake River drains the eastern and southern flanks of the Teton Range.

#### Climate

The Lost River, Lemhi, and Bitterroot Ranges receive most of their moisture from the northern Pacific Ocean and act as an effective barrier to movement of most of the winter airmasses passing to the south out of Canada over the ESRP. This barrier results in low precipitation on the ESRP (Clawson and others, 1989). Additionally, the mountains bordering the ESRP channel the prevailing west winds into a southwesterly flow. Occasionally, however, a cold, Arctic airmass spills over the Continental Divide with a northeasterly flow over the ESRP.

The Yellowstone Plateau and the Island Park Caldera receive 75 percent of their annual precipitation from winter snow that comes from airmasses originating in the Northern Pacific that generally travel up the ESRP (Kharaka and others, 2002). The region also receives Arctic airmasses in the winter and occasional storms originating in the Gulf of Mexico. In the summer months, much of the precipitation comes from warm, moist airmasses originating in the Gulf of Mexico (Dirks and Martner, 1982).

#### Previous Investigations

In 1999, snow cores and summer precipitation in the eastern Island Park region of Idaho were collected and analyzed for stable isotope ratios (Benjamin, 2000). Additionally, precipitation, ground-water, and surface-water stable isotope data from eastern Idaho and western YSNP were compiled from reports by Rightmire and Lewis (1987); Wood and Low (1988); Rye and Truesdell (1993); Bartholomay and others (1994, 1995, 1996); Mann and Low (1994); Ott and others (1994); Knobel and others (1999); Busenberg and others (2000); and Coplen and Kendall (2000). Compilation of regional stable isotope data showed the paucity of existing precipitation data

and underscored the need to collect these data to establish a LMWL for the study area.

A recently published report by Kharaka and others (2002) provides  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  values for 40 snow samples in the YSNP region that define a well-constrained LMWL for YSNP. That line is described by:

$$\delta^2\text{H} = 8.2 \delta^{18}\text{O} + 14.7 \text{ permil VSMOW.} \quad (3)$$

Arctic, Pacific, and Gulf of Mexico airmasses influence weather patterns in the study area, so it is likely that stable isotope ratios for precipitation in the region will reflect the different origins of airmasses and be fairly complex. In the western YSNP/eastern Idaho region, both winter and summer precipitation becomes increasingly lighter in a south-to-north direction in terms of  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  (Rye and Truesdell, 1993; Benjamin, 2000). This south-to-north depletion in stable isotope values probably is the result of a northeasterly Pacific storm trajectory that travels up the ESRP. As these storms travel, precipitation from a storm becomes progressively lighter in isotopic composition. However, it is possible that in a winter where Arctic rather than Pacific airmasses predominate, the south-to-north depletion of stable isotope values would not be seen. It is likely that the climatic effects resulting from the presence of El Niño or La Niña would affect stable isotope values in precipitation (Benjamin, 2000). Precipitation data from Idaho Falls and the upper Henrys Fork Basin show strong seasonal variability in stable isotope values. Summer precipitation is greatly enriched in heavier isotopes relative to winter precipitation; in the Henrys Fork Basin,  $\delta^2\text{H}$  values range from -141 permil (winter) to -55 permil (summer).

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## METHODS

### Sample Collection

Over the period from 1999 to 2001, precipitation samples consisting of snow cores and rainwater were collected from 25 sites located in basins tributary to the Snake River and the ESRPA and in basins tributary to the Madison River in Montana (fig. 2 and table 1). Seven sampling events over this period yielded 75 complete sets of analytical results (including 3 quality-assurance replicates) for samples submitted for laboratory analyses of  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$ .

The data used for this study (table 2) consists of results for 14 samples from 11 sites previously published by Benjamin (2000); results from 2 sites (Copper Basin and Fish Pole Lake) that were sampled only in April 2000 and 2 sites (Teton Pass and State Line) that were sampled only in February 2001; and results from 10 sites that were sampled two or more times during 2000–01.

Sample-collection sites were primarily NRCS snow-telemetry (SNOTEL) and snow-course sites and USGS streamflow-gaging sites. Cores were collected at NRCS SNOTEL and snow-course sites in cooperation with NRCS personnel. Precipitation samplers were positioned at USGS streamflow-gaging stations and at selected NRCS SNOTEL and snow-course sites.

The focus of the 1999 sample collection (Benjamin, 2000) included in this current study was to capture the average isotopic ratio for the snowpack in early spring (when thickness of the snowpack is at its peak), and to capture the average isotopic ratio for summer precipitation collected during July through September. The 2000–01 sample collection was conducted to further refine the expected seasonal variation of the isotopic ratios. Samples were collected in April to represent peak snowpack, in February to represent mid-winter snowpack, in July to represent late spring and early summer precipitation, and in October to represent late summer and early fall precipitation. Two samples collected from sites visited only in April 2000 (Copper Basin and Fish Pole Lake) and two samples collected from sites visited only in February 2001 (Teton Pass and State Line) were opportunity samples collected by NRCS for use in this study.

### Quality Assurance

Samples of snow cores and precipitation were collected and analyzed for  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  using standard methods (Friedman and Norton, 1970; Coplen and others, 1991). SNOTEL sites were reached by snowmobile, and a Mt. Rose snow corer was used for sample collection. The volume and mass of snow cores were recorded and snow water content was determined, and the entire volume of each sample was transferred to a plastic bag and sealed. After melting, the samples were well mixed and immediately transferred to 50-mL

glass containers and sealed with polyseal caps. Rainwater-precipitation collectors consisted of a 1.9-L glass container with an attached funnel sealed to the top. A layer of mineral oil 0.5 cm thick was added to the container to prevent sampling error as a result of evaporation. The precipitation collectors were installed at the sites in April and samples were collected in July and October. A syringe was used to transfer samples to 50-mL glass containers with polyseal caps.

The USGS Stable Isotope Laboratories in Reston, Virginia, and in Menlo Park, California, used a carbon dioxide equilibrium technique (at 30 °C) (Epstein and Mayeda, 1953) to separate the oxygen from hydrogen in the submitted samples, and used mass spectroscopy to determine the ratio of stable oxygen isotopes ( $^{18}\text{O}/^{16}\text{O}$ ). The ratio of stable hydrogen isotopes ( $^2\text{H}/^1\text{H}$ ) is measured by first using a hydrogen gas equilibrium procedure (at 30 °C) (Coplen and others, 1991) to separate the hydrogen, and then mass spectroscopy to determine the ratio of stable hydrogen isotopes. Analytical results are reported as  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  (permil, relative to VSMOW) with 1-sigma uncertainties of 1 permil for  $\delta^2\text{H}$  and 0.1 permil for  $\delta^{18}\text{O}$ . The laboratories analyze each sample for  $\delta^2\text{H}$  in duplicate and every third sample for  $\delta^{18}\text{O}$  to insure the stated uncertainties are achieved or exceeded.

Replicate samples were analyzed for three sites visited in April 2000. These samples, collected from Big Springs/Lucky Dog, Warm River, and Baker Draw, showed relative percent differences for  $\delta^2\text{H}$  ranging from 0.0 to 0.63 percent and for  $\delta^{18}\text{O}$  ranging from 0.03 to 0.24 percent, which were within the expected uncertainties based on a 95-percent confidence interval. Replicate sample results were not used in the linear regression that defines the LMWL.

## RESULTS OF DELTA HYDROGEN-2 AND OXYGEN-18 ANALYSES

Laboratory results for  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  analyses of 75 samples from 25 sites (table 2) were reviewed by applying a linear-regression model to the complete dataset (excluding the 3 quality-assurance replicates). Deuterium-excess (*d*-excess) was calculated for each of 72 results used in the linear-regression model to identify data outliers with the use of Chauvenet's Criterion (Taylor, 1997) and to compare with the normally expected global range of *d*-excess values. The combination of these review tools helped to identify samples that may have been impacted by secondary processes such as partial evaporation of the sample during storage in the rain gage, or by site-specific environmental conditions at the sample location.

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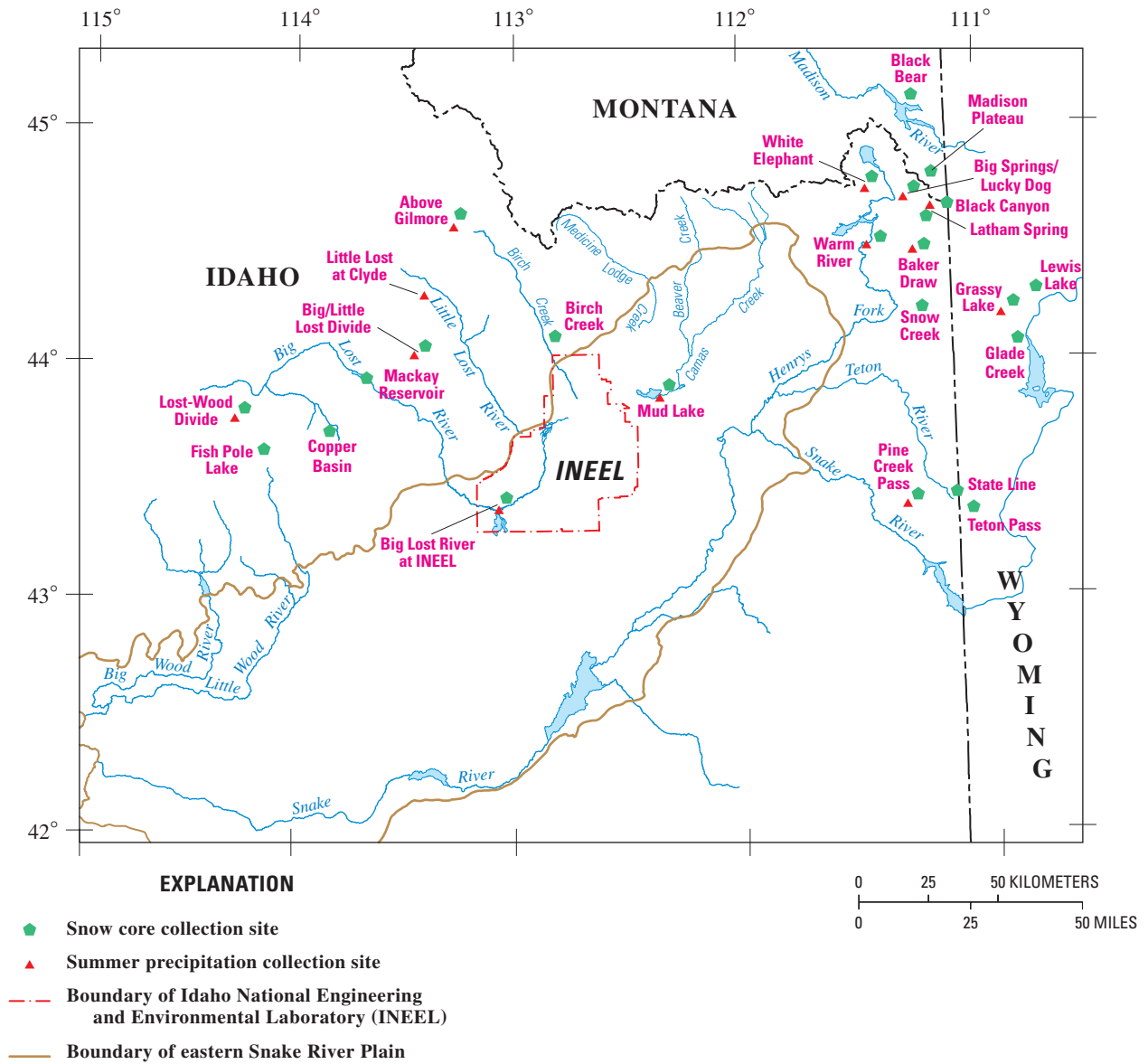


Figure 2. Location of snow core and summer precipitation collection sites, Idaho National Engineering and Environmental Laboratory, and the eastern Snake River Plain.

Table 1. Sampling site elevations and drainage basins in southeastern Idaho, western Wyoming, and south-central Montana, 1999-2001.

[See figure 2 for location of sampling sites. **Bold** indicates site has additional data from Benjamin (2000). Elevation, m above sea level. Drainage basin, drainage basin represented by sample. Abbreviations: BC, Birch Creek; BLR, Big Lost River; BWR, Big Wood River; CC, Camas Creek; HF, Henrys Fork, Snake River; INEEL, Idaho National Engineering and Environmental Laboratory; LLR, Little Lost River; LWR, Little Wood River; MR, Madison River; SF, Upper Snake River (South Fork); SNOTEL, snow telemetry; USGS, U.S. Geological Survey]

Sampling site	Elevation	Drainage basin
Lost-Wood Divide	2,438	BLR/BWR
Mackay Reservoir (USGS streamflow-gaging station)	1,828	BLR
Big Lost River at INEEL (USGS streamflow-gaging station below INEEL diversion)	1,590	BLR
Birch Creek (meteorological station)	2,243	BC
Above Gilmore	2,408	Lemhi/BC
Mud Lake (USGS streamflow-gaging station)	1,455	CC/Mud Lake
White Elephant (SNOTEL site)	2,350	HF
<b>Big Springs/Lucky Dog</b>	1,951	HF
<b>Latham Spring</b>	2,316	HF
<b>Grassy Lake</b>	2,214	Lower HF
<b>Warm River</b>	1,780	HF
Big/Little Lost Divide	2,316	BLR/LLR
Little Lost at Clyde	2,195	LLR
Pine Creek Pass (SNOTEL site)	2,042	Teton River
Copper Basin	2,391	BLR
Fish Pole Lake	2,926	LWR
<b>Baker Draw</b>	2,268	HF
<b>Snow Creek</b>	2,133	HF
<b>Black Bear</b>	2,484	MR
Teton Pass	2,359	Teton River
State Line	2,030	Teton River
<b>Madison Plateau</b>	2,362	MR
<b>Black Canyon</b>	2,426	HF
<b>Glade Creek</b>	2,146	SF
<b>Lewis Lake</b>	2,393	SF



## Linear-Regression Model

Residuals from the regression model (the difference between the observed  $\delta^2\text{H}$  for a given  $\delta^{18}\text{O}$  and the  $\delta^2\text{H}$  predicted for that  $\delta^{18}\text{O}$  by the model) were examined to identify results that fell outside the 95-percent confidence interval for the dataset. Results for three sites—the July 2000 precipitation samples from the Big Lost River at INEEL and from the Big/Little Lost Divide, and the February 2001 snow core from Grassy Lake—were outside the acceptable confidence interval.

## Calculation of Deuterium-Excess

Deuterium-excess is a measure of the relative proportions of  $^2\text{H}$  and  $^{18}\text{O}$  in the sample. Deuterium-excess, defined by  $d$  (in permil) =  $\delta^2\text{H} - 8 \delta^{18}\text{O}$  (Dansgaard, 1964), can be thought of as an index of deviation from the GMWL, which has a  $d$ -excess value of 10 permil. Review of  $d$ -excess values can be a useful diagnostic tool for  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  results because  $d$ -excess can be correlated with conditions at the source area for the water vapor in an airmass and the nature of the airmass prior to the moisture falling as rain or snow (Clark and Fritz, 1997; Froehlich and others, 2002). Partial evaporation of the sample, either as the precipitation falls from the cloud or during storage in the rain gage during warm and dry conditions, can result in low or even negative  $d$ -excess values. Low temperature and lower humidity in the source area for the water vapor that the precipitation is derived from yield large  $d$ -excess values. On a global scale,  $d$ -excess values range from about -2 permil to about 10 to 15 permil (Froehlich and others, 2002). For the northern hemisphere,  $d$ -excess values tend to be largest in December and January and smallest in June and July (Kreutz and others, 2003).

The two tests that were applied to review the  $d$ -excess values both confirmed that the three results identified by examination of the residuals from the linear-regression model required closer review. In addition, the February 2001 snow core from Mud Lake was identified for further review by the  $d$ -excess evaluation.

Expected  $d$ -excess range comparison.—Review of  $d$ -excess values for the 72 primary pairs of  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  results identified 4 values that exceeded the expected range. The July 2000 precipitation values for samples from Big Lost River at INEEL and Big/Little Lost Divide, and the February 2001 snow cores from Grassy Lake and Mud Lake ( $d$ -excess values of -13, -20, 33, and 22 permil, respectively) were outside the expected range of -2 to 15 permil. The first three sites are the same ones identified by the linear-regression model.

Chauvenet's Criterion.—Chauvenet's Criterion (Taylor, 1997, p. 166–169) was used as a test for identifying outliers. This test also was applied to the calculated  $d$ -excess values for all 72 primary pairs of  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  results. Chauvenet's Criterion identified three outlier values; these values were for samples from the same three sites identified by the linear-regression model: the July 2000 precipitation samples from Big

Lost River at INEEL and Big/Little Lost Divide, and the February 2001 snow core from Grassy Lake.

Because Chauvenet's Criterion utilizes the standard-normal probability function, it was assumed that the population of  $d$ -excess values was normally distributed. To insure that the application of Chauvenet's Criterion was appropriate for this dataset, the Shapiro-Wilk W-test (Royston, 1992) was used to test for normality. When applied to the 72 primary data pairs, the W-test indicated that the data were not normally distributed ( $W = 0.8049$ ,  $p = <0.0001$ ). Removing the three values identified as outliers by Chauvenet's Criterion and the one additional value identified by expected  $d$ -excess comparison and repeating the test indicated that the dataset displayed a normal distribution ( $W = 0.9690$ ,  $p = 0.0888$ ) at a 95-percent confidence level.

On the basis of the residual review, the expected range of global  $d$ -excess values, and outlier analysis by Chauvenet's Criterion, these four values were excluded from further regression analysis of all datasets.

Post-depositional modification of the stable isotope signal via melting, evaporation, sublimation, wind erosion, and (or) diffusive mixing of water vapor may be a possible local-scale explanation for the extreme  $d$ -excess values of the four samples excluded from analysis. Research in the Andes Mountains of South America has demonstrated that seasonal changes in snow melting and evaporation rates can amplify seasonal  $\delta^{18}\text{O}$  profiles (Grootes and others, 1989). It was observed that melting of the snow does not affect the  $d$ -excess values but subsequent evaporation may affect  $d$ -excess over diurnal and longer timescales (Clark and Fritz, 1997). Additionally, sublimation during dry winters also may alter  $d$ -excess values. However, none of the slopes of the regression lines for the datasets listed in table 3 approach typical sublimation (evaporation) line slopes (about 5, Clark and Fritz, 1997). It is likely that some type of secondary evaporation processes impacted the four values excluded from analysis.

## DEVELOPMENT OF LOCAL METEORIC WATER LINE

Linear-regression analysis was completed for the final 68 sample results to determine the LMWL for the study area. Regression analysis also was completed for the dataset sorted to investigate seasonal variation. Results and regression parameters are listed in table 3. Regression parameters comprise the correlation coefficient ( $r^2$ ), the standard error for the regression (SE), the probability value (p-value) for the slope of the regression, and the number of samples considered for each analysis. Also included is the 95-percent confidence interval for the slope of the regression line which is identified as  $m$ , and the intercept which is identified as  $b$ . A regression model is a good fit to the data when the correlation coefficient is large (approaching 1.0), the standard error is relatively small in relation to the magnitude of the data, and the p-value is small, approaching less than 0.0001 (Helsel and Hirsch, 1992). The

## 10 Development of a local meteoric water line for southeastern Idaho, western Wyoming, and south-central Montana

confidence in the regression parameters for describing the fit of the regression model to the data tends to increase with the number of samples (larger datasets,  $n$  greater than 30, table 3).

The regression for all precipitation data (snow cores and rain water, minus the four values excluded following review of regression residuals and  $d$ -excess, figure 3) yielded a slope of 7.95 ( $n = 68$ ,  $r^2 = 0.98$ ), near the expected slope of 8.0 for the GMWL. The Y-intercept ( $d$ -excess value) for the regression is 8.09, between the 6 derived for the North American MWL (Yurtsever and Gat, 1981) and the  $d$ -excess of 10 for the GMWL. The data also were sorted by seasons (snow cores, 1999–2001, and precipitation, 1999–2000), and the regression does show the expected seasonal variability (fig. 4), as winter snow plots lighter than summer precipitation. The summer precipitation also shows a much wider range of values, representing different sources of precipitation from differing storm tracks in the summer, as well as differing evaporative conditions. The slope for winter precipitation (bulk snow cores, 1999–2001) is 7.41 and the  $d$ -excess is -2.18 ( $n = 46$ ,  $r^2 = 0.89$ ). The slope for summer precipitation (May–October 1999–2000) is 8.49 and the  $d$ -excess is 14.5 ( $n = 22$ ,  $r^2 = 0.96$ ). The final dataset used to establish a LMWL for this study and the four excluded values are shown in figure 5. Also included for comparison are regression lines with the 95-percent confidence intervals for the final dataset and for the dataset containing the four excluded values.

The meteoric water line specific to a location or region is controlled by local factors specific to the climate of that region, including the oceanic origin of the water vapor and storm tracks characteristic of the region, and secondary evaporation of the precipitation. Such local factors work together to produce the slope and  $d$ -excess specific to that LMWL. The LMWL for the study area should be defined by the final set of all data,

excluding the four  $d$ -excess values outside the expected range. This line is  $\delta^2\text{H} = 7.95 \delta^{18}\text{O} + 8.09$ . The slope of this LMWL is not different from that for the 1976–77 Idaho Falls precipitation data ( $\delta^2\text{H} = 7.94 \delta^{18}\text{O} + 3.12$ ) (table 3), but the Y-intercept is different, and this may be a reflection of the more regional nature of the dataset used in this study. The LMWL developed in this study and the one for YSNP developed by Kharaka and others (2002) ( $\delta^2\text{H} = 8.2 \delta^{18}\text{O} + 14.7$ ) also are similar (table 3). The small differences in  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  values used to develop the lines result from differences in the amount of evaporation that takes place in the two study areas. The YSNP line (developed from values derived from 40 snow samples) generally reflects less evaporative isotope fractionation than does the line reported in this study (developed from values derived from 68 rain and snow samples).

## SUMMARY AND CONCLUSIONS

Linear-regression analysis of 68 pairs of  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  values yielded a Local Meteoric Water Line (LMWL) defined by the equation  $\delta^2\text{H} = 7.95 \delta^{18}\text{O} + 8.09$ . The slope of this LMWL is not different from that for the 1976–77 Idaho Falls precipitation data ( $\delta^2\text{H} = 7.94 \delta^{18}\text{O} + 3.12$ ), but the Y-intercept is different, and this may be a reflection of the more regional nature of the dataset used in this study. The slope and intercept of the LMWL also are similar to those of the Global Meteoric Water Line ( $\delta^2\text{H} = 8 \delta^{18}\text{O} + 10$ ) and the YSNP LMWL ( $\delta^2\text{H} = 8.2 \delta^{18}\text{O} + 14.7$ ). This likely results from the averaging effects of storms originating in different source areas (Gulf of Mexico, Pacific Ocean, and Arctic Ocean) and from small differences in evaporative isotope fractionation of H and O, respectively.



Table 3. Linear regression results for snow core and precipitation samples, southeastern Idaho, western Wyoming, and south-central Montana, 1999–2001, and other historical data.

[Symbols:  $\delta^2\text{H}$ , delta notation for stable hydrogen isotopes in permil relative to Vienna Standard Mean Ocean Water (VSMOW);  $\delta^{18}\text{O}$ , delta notation for stable oxygen isotopes in permil relative to VSMOW]

Dataset	Linear regression parameters for $\delta^2\text{H} = m\delta^{18}\text{O} + b$						
	m (Slope)	b (Y-intercept)	r <sup>2</sup> (Correlation coefficient)	SE (Standard error for regression)	95-percent Confidence interval of m	p-value for the slope of the regression)	n (number of samples)
All samples analyzed for this study (minus the 4 excluded values, see figure 3 and text)							
<b>Season (see figure 4)</b>							
Bulk snow cores, 1999–2001	7.41	-2.18	0.89	3.09	6.63 to 8.20	Less than 0.0001	46
Precipitation 1999–2000	8.49	14.5	0.96	4.29	7.68 to 9.30	Less than 0.0001	22
<b>Other historical data</b>							
Idaho Falls precipitation, 1976–77 <sup>1</sup>	7.94	3.12	0.99	5.26	7.64 to 8.25	Less than 0.0001	40
Global Meteoric Water Line <sup>2</sup>	8	10					
North American Meteoric Water Line <sup>3</sup>	8	6					
Yellowstone Park Meteoric Water Line <sup>4</sup>	8.2	14.7					

<sup>1</sup> Rightmire and Lewis (1987).

<sup>2</sup> Craig (1961).

<sup>3</sup> Yurtsever and Gat (1981).

<sup>4</sup> Kharaka and others (2002).

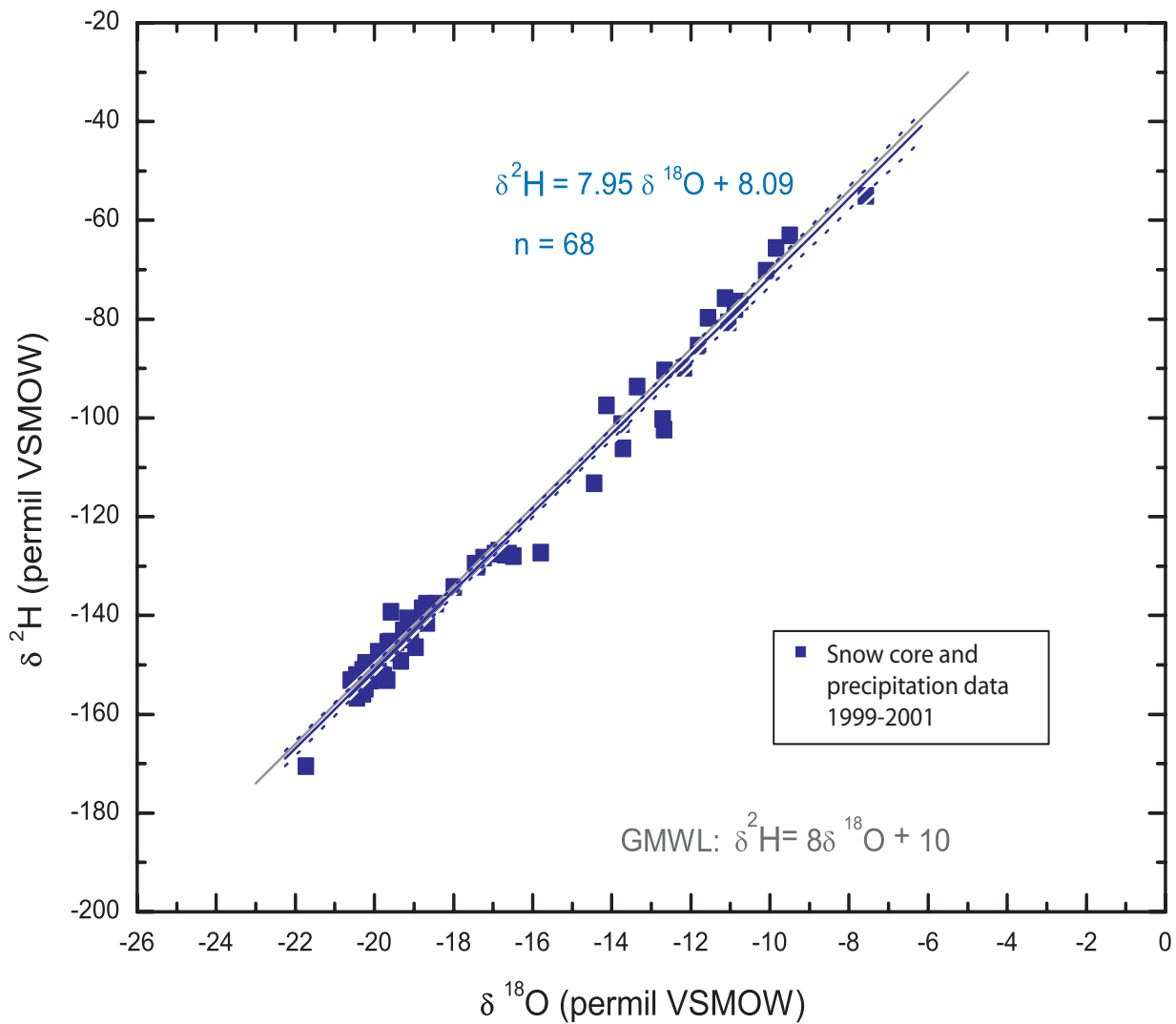


Figure 3. Local Meteoric Water Line for southeastern Idaho, western Wyoming, and south-central Montana [ $\delta^2\text{H}$ , delta notation for stable hydrogen isotopes in per mil relative to Vienna Standard Mean Ocean Water (VSMOW);  $\delta^{18}\text{O}$ , delta notation for stable oxygen isotopes; GMWL, Global Meteoric Water Line]

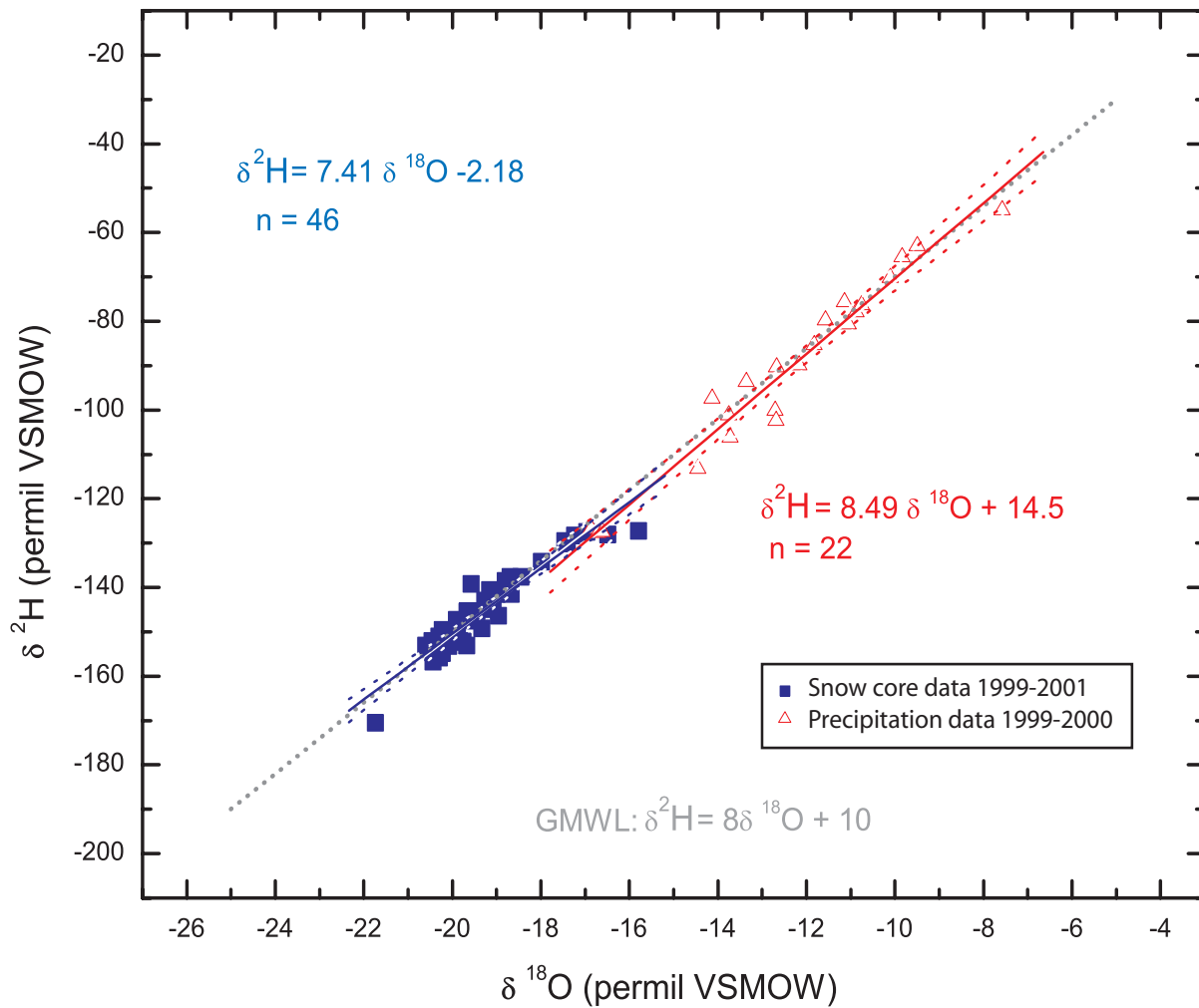


Figure 4. Seasonal variation in Local Meteoric Water Line for southeastern Idaho, western Wyoming, and south-central Montana [ $\delta^2\text{H}$ , delta notation for stable hydrogen isotopes in permil relative to Vienna Standard Mean Ocean Water (VSMOW);  $\delta^{18}\text{O}$ , delta notation for stable oxygen isotopes; GMWL, Global Meteoric Water Line]

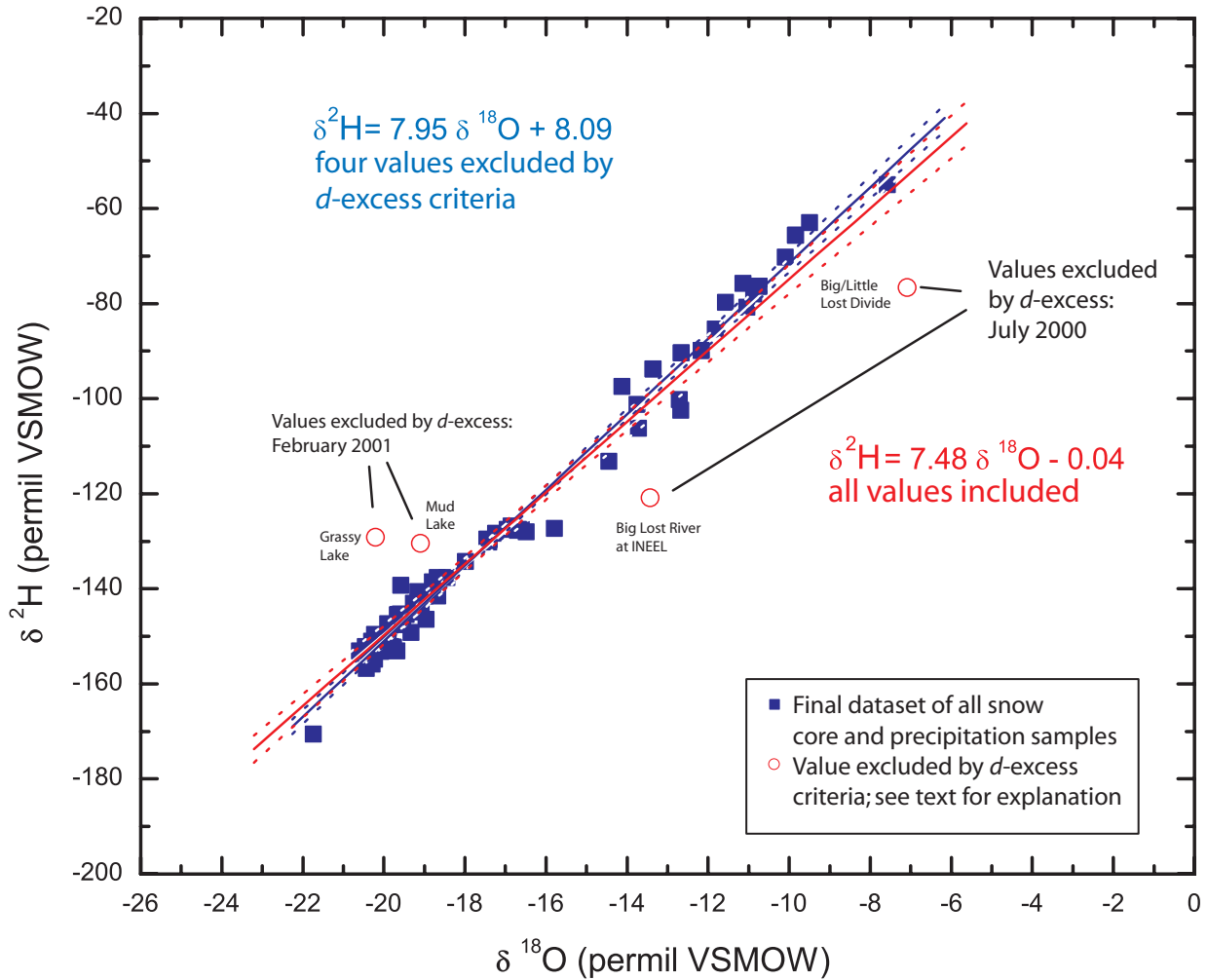


Figure 5. Final dataset used to establish a Local Meteoric Water Line for southeastern Idaho, western Wyoming, and south-central Montana [Values excluded by extreme deuterium-excess criteria also shown;  $\delta^2\text{H}$ , delta notation for stable hydrogen isotopes in permil relative to Vienna Standard Mean Ocean Water (VSMOW);  $\delta^{18}\text{O}$ , delta notation for stable oxygen isotopes]

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