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Revised Quaternary glacial succession and post-LGM recession, southern Wind River Range, Wyoming, USA.

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Abstract

We present here a more complete cosmogenic chronology of Pleistocene glacial deposits for the Wind River Range, Wyoming, USA. Fifty-one new and thirty-nine re-calculated ^{10}Be and ^{26}Al exposure ages from Sinks and North Fork canyons, Stough Basin, Cirque of the Towers and the Temple Lake valley allow us to more tightly constrain the timing and sequence of glacial alloformations in the southern portion of the range.

Moraines, diamicts and bedrock exposures here have previously been correlated with as many as five Pleistocene and four Holocene glacial events. Exposure ages from Pleistocene alloformations associated with trunk glaciers in Sinks Canyon and North Fork Canyon generally confirm earlier age estimates. Cosmogenic radionuclide (CRN, ^{10}Be and ^{26}Al) ages from moraines and striated bedrock surfaces previously mapped as Pinedale correspond to MIS2, while boulder exposure ages from moraines mapped as Bull Lake correspond generally to MIS5-MIS6. Geomorphic data from a moraine previously mapped as Younger pre-Sacagawea Ridge appears to correspond most

closely to the Sacagawea Ridge glacial episode (MIS-16), but the uncertainty of a single ^{10}Be exposure age suggests the unit could be as young as MIS-10 or as old as MIS-18. Boulders from a diamict on Table Mountain previously reported as Older pre-Sacagawea Ridge yield two ^{10}Be exposure ages that suggest the presence of Early Pleistocene glacial activity here possibly older than ~1-2 Ma (>MIS-30).

Bedrock exposure ages within Sinks Canyon suggest the Pinedale valley glacier had retreated from the floor of Sinks Canyon to above PopoAgie Falls by ~15.3 ka. Cirque glaciers in Stough Basin appear to have retreated behind their riegels by ~16 ka, which suggests the cirque glaciers were decoupling across their riegels from the valley glaciers below at this time, prior to their readvance to form Lateglacial moraines.

New ^{10}Be boulder exposure ages from moraines previously correlated to the Temple Lake and Alice Lake allostratigraphic units in the cirques of Stough Basin and Cirque of the Towers show general equivalence to the stadial event just prior to the onset of the Bølling interstadial (17.5-14.7 ka) and to the Intra-Allerød Cold Period-Younger Dryas stadial phase (13.9-11.7 ka), respectively. From this evidence, the Temple Lake Alloformation of the Wind River Mountains now should correspond to the INTIMATE GS-2.1a (Oldest Dryas) stadial event while the Alice Lake Alloformation should correspond to the INTIMATE GS-2 stadial (IACP-Younger Dryas). Thus, we consider that evidence no longer exists for early- to mid-Holocene glacial events in the southern Wind River Range.

1. Introduction

The U.S. Rocky Mountains contain numerous ranges with records of multiple Pleistocene glacial episodes (Richmond, 1986; Dahms, 2004b; Locke, 1990; Locke and Smith, 2004; Osborn and Gerloff, 1997; Pierce, 2004; Pierce et al., 2018). The Greater Yellowstone Ecosystem of Wyoming-Montana-Idaho plays a central role in our understanding of North American alpine glacial history as it contains many of the type localities used for our present understanding of the Pleistocene-Holocene glacial succession in this region of the Rockies (e.g., Richmond, 1986; Dahms, 2004b; Dahms et al., 2010; Pierce, 2004). The Wind River Range (WRR) occupies the southern-most portion of the Greater Yellowstone Geoecosystem and, along with Yellowstone Park itself, is the focus of much past and continuing research into the regional Pleistocene glacial succession (Blackwelder, 1915; Richmond, 1948, 1964, 1965, 1973, 1986; Richmond and Murphy, 1965, 1989; Murphy and Richmond, 1965; Mears, 1974, Dahms, 2004a, b; Dahms et al., 2010).

1.1 Pleistocene Succession of the WRR

Glaciers in the Wind River Mountains have occupied all of the range's major alpine valleys. Using the seminal works of Blackwelder (1915) and Love (Mears, 1974), Richmond (1965, 1986 and references therein) presented morphostratigraphic evidence from the Bull Lake Type Area (BLTA) that identified a series of moraine, outwash and lake deposits at Cedar Ridge corresponding to five purported early-to-late Pleistocene glacial periods [from youngest: Pinedale - Bull Lake - Sacagawea Ridge - Cedar Ridge - Washakie Point]. Hall and Jaworowski's (1999) reevaluation of the Cedar Ridge section showed that all of the Pleistocene allostratigraphic units (NACSN, 1983) above the Tertiary beds at Cedar Ridge should be correlated to Sacagawea Ridge-and-younger deposits and are paleomagnetic-normal (no older than the Gauss-Matuyama boundary of 781 ka). Thus, no evidence exists at this locality for Richmond's purported Cedar Ridge and Washakie Point deposits. Likewise, recent ^{10}Be and ^{36}Cl exposure age-analyses from moraine boulders at the BLTA (Hall and Jaworowski, 1999; Chadwick et al., 1997) also found no evidence for pre-Sacagawea Ridge units. Thus, since 1999, the oldest two of Richmond's three 'pre-Bull Lake' units are no longer viable allostratigraphic units in the WRR, and that only the Sacagawea Ridge, Bull Lake, and Pinedale remain as widely recognized units.

Dahms (2004a) used morphostratigraphy and soil development at Sinks Canyon to identify a succession of allostratigraphic units (moraines) corresponding to the Pinedale (MIS2; Cohen and Gibbard, 2011), Early Wisconsin (MIS4), Bull Lake (MIS6), Sacagawea Ridge (MIS16?) glaciations as well as two stratigraphically older diamictons above/outside the canyon that suggested that two older (undated) glacial advances were represented here. These were provisionally termed Older and Younger pre-Sacagawea Ridge (Dahms, 2004a).

The previous model for the Lateglacial/Holocene (post-LGM) succession of the WRR (Dahms et al., 2010) was based on cumulative relative and numeric age data gathered by numerous workers from alpine valleys along the range (Holmes and Moss, 1955; Currey, 1974; Dahms, 2002; Gosse et al., 1995a, b, 1999; Mears, 1974; Miller and Birkeland, 1974; Mahaney, 1978, 1984a, b; Zielinski and Davis, 1987). The main points of contention in this work have been (a) the age of those deposits previously associated with the Younger Dryas (YD) and (b) the number and age(s) of post-YD events preserved here. Early interpretations of the post-LGM succession focused chiefly on the age of the type Temple Lake moraine in the Temple Lake Valley. Hack (1943) and Moss (1949, 1951; Holmes and Moss, 1955) identified deposits corresponding to two late Pleistocene – Holocene glacial advances here. Their work identified the Type Temple Lake moraine as a pre-Altithermal unit and younger moraines corresponding to the 'Little Glaciation' (Little Ice Age). Richmond (1965) later revised the interpretations of Holmes and Moss in the Temple Lake valley, suggesting that two Temple Lake moraines were preserved here *("a"* and *"b"*) that represented the oldest two of three neoglacial (post-Altithermal) advances. Richmond also revised the name of the Little Glaciation to Gannett Peak (Richmond, 1965; Benedict, 1968; Birkeland et al., 1971). Miller and Birkeland (1974) later re-interpreted these deposits, using significant differences in moraine and boulder weathering characteristics and soil development to suggest four YD-to-Holocene glacial events are preserved here [Temple Lake, Early Neoglacial, Audubon equivalent (Benedict, 1973; Miller and Birkeland, 1974), Gannett Peak]. Most recently, Dahms (2002) and Dahms et al. (2010) presented a revised post-LGM stratigraphy for the WRR that essentially mirrored Birkeland and Miller's correlations with suggested ages: Gannett Peak (LIA), Black Joe (1-2 ka), Alice Lake (~5500-4000 yr), Temple Lake (YD-equivalent).

In this paper, we use a combination of new and recalculated ^{10}Be and ^{26}Al exposure ages from successions of moraines, diamictons, and bedrock surfaces previously described by Dahms (2002, 2004a) and Fabel et al. (2004) from Table Mountain, Sinks Canyon-Stough Basin, and North Fork Canyon-Cirque of the Towers to more tightly constrain the Pleistocene glacial succession for the southern WRR (Fig. 2). As the WRR is the type locality for most of the Rocky Mountain glacial sequence, an updated chronology here adds to a more complete understanding of the alpine glacial succession in North America. Additionally, we present evidence for rates of ice retreat along the Middle and North Forks of the PopoAgie (Po-po'-zhuh) Basin from the maximum positions of Pinedale ice (MIS2) in Sinks and North Fork canyons at/near the Last Glacial Maximum (LGM) to the Lateglacial positions of the cirque glaciers as represented by moraines in Stough Basin and Cirque of the Towers.

2. Study Area

The Wind River Range (WRR) is located in the Middle Rocky Mountains of west-central Wyoming, with the PopoAgie River basin on the range's southeastern flank (Fig. 1). Table Mountain, Sinks Canyon and Stough Basin are parts of the Middle PopoAgie Basin while North Fork Canyon and the Cirque of the Towers occupy most of the basin of the North Fork of the PopoAgie (Fig. 2). Table Mountain and the mouth of Sinks Canyon are located ~15 km southwest of the city of Lander, Wyoming. Sinks Canyon was the single outlet for the trunk glacier in the Middle PopoAgie Basin and is the most southerly of the four major canyons along the eastern slope of the WRR (Fig. 1) where Pleistocene glacial deposits previously were described (Richmond, 1957, 1986; Richmond and Murphy, 1965, 1989; Murphy and Richmond, 1965; Shroba, 1989; Chadwick et al., 1997; Phillips et al., 1997; Applegarth and Dahms, 2001; Dahms, 2004a). The mouth of North Fork Canyon lies ~13 km northwest of Lander (Fig. 2). North Fork Canyon was the single outlet for the North Fork Basin trunk glacier. Post-LGM glacial deposits were previously reported from Stough Basin and Cirque of the Towers by Dahms and his colleagues (Dahms, 2002; Dahms et al., 2010). Glacial deposits have not previously been described from the North Fork Canyon, downvalley from Cirque of the Towers.

Bedrock of the areas sampled for this study is Archaean granite and granodiorite of the Louis Lake Formation (Love and Christianson, 1985; Frost et al., 2000). Although Sinks and North Fork canyons are carved into a nearly complete section of the Paleozoic limestones, dolomites, and sandstones described for this region of Wyoming (Love et al., 1992), only granitic boulders and bedrock exposures were sampled for this study.

3. Materials and methods

We combined recalculated ^{10}Be and ^{26}Al exposure age-data with newly-generated ^{10}Be ages to obtain the best possible insight into the timing of the glacial advances and of the post-LGM ice retreat in our study region. We acquired nineteen new samples from boulders on previously identified moraines (Dahms, 2004a) associated with the Sinks Canyon trunk glacier (Pinedale, Bull Lake, Sacagawea Ridge and pre-Sacagawea Ridge) and a previously unreported moraine at the mouth of North Fork Canyon (Pine Bar Ranch). In the upper portion of Sinks Canyon, we re-calculated eighteen ages from two previously-reported valley-side transects of polished-striated bedrock (Fabel et al., 2004). We also present fifteen new ages from two cross-basin bedrock transects in Stough Basin.

We report ten new ^{10}Be exposure ages from boulders on alpine moraines in Helen cirque (Stough Basin) and in Cirque of the Towers that Dahms (2002; et al., 2010) previously correlated to the Younger Dryas and 'Neoglacial' (Temple Lake and Alice Lake Alloformations). We include in our interpretations fourteen recalculated ^{10}Be exposure ages previously developed by Marcott (2011) from boulders on the Alice Lake-age and Temple Lake-age moraines in Bigfoot cirque (Fig. 6 in Dahms et al., 2010) as well as eight boulders from the Type Temple Lake moraine in the Temple Lake Valley (Marcott, 2011; locations S5a-S6b of Fig. 5C in Dahms et al, 2010). In order to compare our interpretations with other regions in North America, we have recalculated surface exposure-ages published prior to 2011.

3.1 ^{10}Be and ^{26}Al surface exposure ages

In order to obtain the most reliable exposure ages, we used commonly-accepted methods for sampling moraine boulders (e.g., Gosse and Phillips, 2001; Masarik and Wieler, 2003). We chose the largest available boulders with glacial polish and/or striations protruding more than 1m from stable moraine ridges to avoid post-depositional tilting. We sampled boulders with relatively flat tops to avoid edge effects. We also sampled exposures along two valley cross-sections in both Sinks Canyon (Fabel et al., 2004) and Stough Basin that showed clear evidence of glacial polish and/or striae.

The position (latitude/longitude and altitude) of each sample site was recorded with GPS and verified with a topographic map as both have an influence on the amount of cosmic radiation. We measured the dip of the boulder surface, the direction of the dip and the topographic shielding to correct for the geometry of individual boulders and the effect(s) of topographic shielding by surrounding mountains. We sampled only the uppermost 1-3 cm of boulders and 1-5 cm of the bedrock transects and documented sample thickness to account for the attenuation of cosmic rays with depth within the rock material.

For ^{10}Be analyses the rock samples were pre-treated following the procedures of Kohl and Nishiizumi (1992) and Ivy-Ochs (1996). Samples were crushed and sieved and the quartz isolated by treating the 0.25mm–0.6 mm fraction with *aqua regia* to destroy organic contaminations and any calcareous components. After a 1h-treatment with 0.4% HF, we used a floatation system to physically separate feldspar and mica components from quartz. Remnant feldspars and micas were removed by repeated 4%HF leaching. Once pure quartz was obtained, we added a ^{9}Be-carrier solution and dissolved the samples in 40%HF. Isotopic beryllium was isolated using anion and cation exchange columns followed by selective pH precipitation techniques (von Blanckenburg et al., 1996). The Be hydroxides were precipitated, dried, and calcinated to BeO at 850°C. The $^{10}Be/^{9}Be$ ratios were measured at two different accelerator mass spectrometry facilities. New samples in Table 1 were analysed at the ETH Laboratory of Ion Beam Physics' Accelerator Mass Spectrometry (AMS) facility using the ^{10}Be standard S2007N with a nominal value of $^{10}Be/^{9}Be = 28.1 \times 10^{-12}$ (Christl et al., 2013; Kubik and Christl, 2010). S2007N has been calibrated to the ^{10}Be standard ICN 01-5-1 of K. Nishiizumi and has a nominal $^{10}Be/^{9}Be$ value of 2.709×10^{-11}(Nishiizumi et al., 2007). The 1σ error of S2007N is 2.7% (Christl et al., 2013). New ^{10}Be and ^{26}Al data in Table 2 were determined at PRIME Lab, Purdue University between 1997 and 1999 and normalised to NIST SRM4325 with $^{10}Be/^{9}Be$ 2.68×10^{-11}, and Z92-0222 with $^{26}Al/^{27}Al$ 4.11×10^{-11}. The original PRIME Lab $^{10}Be/^{9}Be$ results have been converted to be directly comparable to the above $^{10}Be/^{9}Be$ standard value of 2.709×10^{-11}(Nishiizumi et al., 2007). Measured $^{10}Be/^{9}Be$ ratios were corrected for ^{10}Be contributed by the Be-carrier determined from process blanks ($^{10}Be/^{9}Be$ of 3.0×10^{-15} for both AMS laboratories). No correction was required for $^{26}Al/^{27}A/$ measurements. Stable Al concentrations in aliquots of the dissolved quartz were determined by flame atomic absorption spectrophotometry (AAS), using the method of standard additions. Little to no matrix effect was observed, and [Al] measurements were reproducible to 2%. ^{10}Be and ^{26}Al AMS data and concentrations for the boulder and bedrock samples are reported in Table S1 and S2, respectively.

All exposure ages reported here are calculated using CRONUS-Earth version 2.3 (http://hess.ess.washington.edu/math/) with the default production rates (4.01 ^{10}Be atoms/gram SiO_2/year, 27.07 ^{26}Al atoms/gram SiO_2/year; Borchers et al., 2016) and half-lives (^{10}Be half-life of 1.387±0.012 Ma (Chmeleff et al. 2010; Korschinek et al., 2010) and ^{26}Al half-life of 0.705±0.018 Ma

(Nishiizumi 2004). The production rate was scaled for latitude, Longitude and altitude using the time-dependent Lm scaling scheme (Lal, 1991; Stone, 2000). We corrected for sample thickness assuming an exponential depth profile (Brown et al., 1992) with an effective radiation attenuation length of 160 g cm^{-2} (Gosse and Phillips, 2001) and a rock density of 2.7 g cm^{-3}. Following Marcott (2011) we assumed a rock erosion rate of 0mm/ky for samples from those boulders ($\leq$LGM) that exhibited glacial polish and striae. We used erosion rates of up to 2mm/ky for boulders on older Pleistocene deposits that exhibited progressively greater degrees of weathering with presumed age according to their geomorphic relations. Although this method is rather subjective, these relations appear to be useful. Note that the ages that we derive from boulders on Pinedale (0 mm/ky erosion rate) vs Bull Lake (2 mm/ky erosion rate) correspond well to ages of Pinedale and Bull Lake moraines elsewhere in the WRR. We applied no correction for snow. Surface exposure ages with one-sigma uncertainties for the boulder and bedrock samples are reported in Tables 1 and 2, respectively.

4. Results and Discussion

4.1 Table Mountain (Early Pleistocene)

Dahms (2004a) described the diamict on Table Mountain (Figs. 2, 3) and interpreted its age as 'pre-Sacagawea Ridge' (Early Pleistocene) on the basis of its geomorphic/stratigraphic position and the soil weathering characteristics. ^{10}Be ages reported here from eight boulders at several positions across Table Mountain range from >2000 to ca. 150 ka (Fig. 3, Table 1). The variability of the exposure ages suggests two immediate interpretations: that (1) this material was deposited in Early Pleistocene time and that (2) either many of the boulders were buried for much of their histories and have more recently become exposed at the surface or those boulders with relatively young exposure ages have undergone significant erosion since their deposition, or both.

The depths of original moraine matrix removed from above the once-buried boulders is unknown, but it is apparently significant. Moraine erosion of 3-4 mm/ky was needed to bring the ^{10}Be ages of 1.5 m-diameter boulders at Dinwoody Lake, into concordance with the age of a 3 m-diameter boulder, so that "… in 650 ka, an erosion rate of 0.35 cm ka (sic) would remove more than 2 m of till from the crest" (Gosse et al., 2003). A similar calculation south of Dinwoody Lake on the Bull

Lake moraines at the BLTA suggested these moraine crests had lost more than 1.4 m of till in ~140 ka. Using these erosion estimates and our estimated ages of >1-2 Ma for the two oldest boulders, it is possible that as much as 14 m of material has been removed from some portions of the Table Mountain diamicton.

Boulder heights here do not appear to follow the common assumption that higher boulders exhibit more dependable exposure ages (e.g., Gosse et al., 1995a). Heyman et al. (2016) show a generally positive correlation between taller boulders and exposure age groups, but also note that a dominant fraction of the groups still have scattered exposure ages. In this study, however, the two oldest Table Mountain boulders (TM-1, 6) are less than three meters tall, while the tallest boulders (ET-1, 2) exhibit comparably young ages. While Gosse et al. (2003) reported no relation between boulder height and ^{10}Be age on Bull Lake moraines at Fremont Lake (western slope WRR) they reported a positive correlation on the Sacagawea Ridge moraine near Dinwoody Lake (eastern slope). There are many possibilities for how boulders erode, either fast or slow, on any specific landform, but the most obvious mechanism for boulder erosion over time at the present location are fire and lightning strikes (Zimmerman et al., 1994; J.C. Gosse personal communication). The high elevation of Table Mountain leaves it extremely exposed to thunderstorm activity and many Lander-area residents have stories of lightning strikes and near-misses here during summer afternoon horse-back rides.

The exposure ages (1-2 Ma) of the oldest two of the Table Mountain boulders (Fig. 3; TM-1, TM-6) suggest that ice extended outside whatever form that Sinks Canyon took during the Early Pleistocene and flowed over the position now occupied by the canyon of Sawmill Creek onto the surface of Table Mountain (Dahms, 2004a; Züst et al., 2014). A re-entrant valley which now separates the eastern from the western half of Table Mountain (Fig. 3) suggests that two separate ice advances might be represented here. It appears that an older ice advance delivered material to the eastern end of Table Mountain near the locations of samples ET-1 and ET-2 (Fig. 3). Additional evidence for glacial ice at this location can be seen just north of boulder ET-2. A series of step-like features here (Fig. 3 dashed orange lines) appear to be the remnants of kame terraces that mark the southern margin of such an ice mass. The presence of two surfaces on separate sides of the re-entrant suggests that either a recessional position of an older glacier or a separate (younger) glacier terminated at the central portion of Table Mountain long enough for meltwater to remove substantial amounts of the older material to form the re-entrant valley.

The position of the Table Mountain diamict over 400 m above the present Middle PopoAgie River also suggests that, during the early Pleistocene, Sinks Canyon had not developed to the extent that it's dimensions could contain all of the ice flowing from the basin's head. No erratic materials have been found further outside Sinks Canyon to the north or south to suggest that ice occupied a route different from the present position of the canyon. Similar boulder/diamict materials, however, are found at similar elevations and positions outside/above the canyon mouths at other locations along the eastern slope of the WRR (Veggian et al., 2010). Although no ages are currently available for these deposits, we suggest the Table Mountain diamict represents a more extensive regional pattern of glacial deposition during the early Pleistocene when few canyons had developed to dimensions that would enable them to constrain the ice volume of their valley glaciers. This situation appears to be analogous to Anderson et al's (2012) concept of 'far-flung moraines' where older moraines often are found many kilometres beyond more recent moraines as an inevitable consequence of glacial erosion over time wherever glacial erosion rates are greater than uplift rates. Thus, the presence of these high elevation moraines/diamicts outside canyon mouths suggests that we should revisit older models of 'pre-canyon' glacial events in this region of the Rocky Mountains (Blackwelder, 1915; Richmond, 1948, 1957; Love, 1977; Mears, 1974; Veggian et al., 2010).

4.2 Sacagawea Ridge (Early Middle Pleistocene)

The age of deposits associated with the Sacagawea Ridge glaciation in the WRR remains poorly constrained (Gosse et al., 2003). Outwash terraces containing Lava Creek B tephra (~650 ka) were earlier identified downstream from and/or correlated with Sacagawea Ridge moraines at Dinwoody Lakes (Richmond and Murphy, 1965; Richmond, 1976). More recently, a new locality of a previously identified Lava Creek B ash deposit has been identified in the gravels of the high terrace at the Lander airport (Anders et al., 2009; Dahms and Egli, 2016; William McIntosh, personal written communication, January 2018). However, few boulder exposure ages have been reported from moraines correlated to the Sacagawea Ridge glaciation. Gosse et al. (2003) obtained time-constant ^{10}Be exposure ages of 145-to-360 ka from four boulders on the Type Sacagawea Ridge moraine at Dinwoody Lake. When erosion rate(s) were considered according to boulder heights, the resulting boulder exposure ages were equivalent to ~650 ka. If a time-dependent scaling were applied to this data, we estimate the ages would be >50 ka younger.

Phillips et al. (1997) developed ^{36}Cl exposure ages of ca. 261-to-99 ka from a suite of six boulders on the Sacagawea Ridge moraine at the BLTA. No additional cosmogenic exposure ages have been derived from moraines mapped as 'Sacagawea Ridge' prior to the current study (Table 1 and below).

A left lateral remnant moraine associated with an isolated field of erratic boulders is located south and east of the mouth of Sinks Canyon ~220 meters below Table Mountain and ~150 meters above the Middle PopoAgie River (Fig. 3). This material apparently represents a younger glaciation than that represented by the diamict on Table Mountain. Dahms (2004a) interpreted this material as 'Younger pre-Sacagawea Ridge' on the basis of its correspondence to the moraine mapped on the south rim of Sinks Canyon (see location of DS-1 in Figs 3 and 4) and the moraine's position relative to deposits within the canyon correlated to the Sacagawea Ridge Alloformation.

Erratic boulders scattered across the dipslope between the PopoAgie River and the lower slopes Table Mountain (yellow dots on Fig. 3) are entirely constrained between the above remnant moraine and the base of Table Mountain. No erratic or till material is located on the bedrock dipslope between the moraine and the river (Dahms, 2004a). This pattern suggests the 'Sacagawea Ridge' glacier split into two ice streams at an outcrop of the Tensleep Sandstone just above the Sinks. One ice stream flowed out over the canyon rim to the east across what is now Sawmill Canyon to below the northwest end of Table Mountain (Fig. 3). This ice stream contained boulder DS-1, which yields an exposure age of 556±188 ka. The ice stream remaining in the canyon continued down and out of the canyon, terminating ~3 km outside the canyon mouth. If much of the Sacagawea Ridge glacier's volume remained constrained within the canyon, then Sinks Canyon had enlarged enough by this time to constrain a larger ice volume than during the earlier advance(s) (Table Mountain). If this unit is indeed Sacagawea Ridge-age, its geomorphic position matches earlier interpretations from the BLTA and from Dinwoody Lakes that the Sacagawea Ridge glaciation was the initial post-canyon event in the WRR (Richmond, 1965, 1986).

More evidence for the association of Sacagawea Ridge ice with the outlet of Sinks Canyon occurs just outside the canyon's mouth. An area just north of the river contains a deposit of mixed Sacagawea Ridge and Bull Lake-age materials that we interpret as stagnant ice debris (moraine/outwash)(Fig. 3; Dahms, 2004a). Here Dahms (2004a) and Dahms et al. (2012) described a deeply weathered soil profile (HR-1 in Fig. 3) with all granitic clasts completely weathered to grus to a depth of over 2 meters. Bull Lake-age materials are admixed with the older materials (HR-2 in Fig. 3; see below) as seen by their soil profiles that are less deeply weathered with fewer, smaller clasts weathered to grus (Dahms, 2004a).

Thus, the evidence to support our interpretation that the deposits Dahms (2004a) earlier correlated to a 'Younger pre-Sacagawea Ridge' glaciation are probably no younger than Sacagawea Ridge includes the following:

(1) The associated materials lie above the Middle Popo Agie River between the higher (older) deposits on Table Mountain and the lower (younger) Bull Lake deposits;

(2) the surface of the outwash terrace containing the Lava Creek ash at Lander airport (Anders et al., 2009; Dahms and Egli, 2016; William McIntosh, personal written communication, January 2018) grades to the lowest margin of the moraine/outwash described (above) outside the canyon mouth;

(3) soil weathering characteristics suggest these are younger deposits than those described for Table Mountain and older than those associated with Bull Lake-age deposits (Dahms, 2004a; Dahms et al., 2012);

(4) our ~556 ka exposure age from boulder DS-1 (Figs. 3, 4; Table 1). This exposure age has a relatively large uncertainty. With the 1δ external uncertainty the age could be between 368 ka and 744 ka. Considering only internal error the uncertainty still places the age from 452 ka to 660 ka. Even with these uncertainties, the age certainly falls between that of the Table Mountain diamict mapped above and the Bull Lake moraines mapped below and allows us to make a reasonable assumption that this unit corresponds to the Sacagawea Ridge Alloformation.

4.3 Bull Lake (Late Middle Pleistocene)

The Bull Lake glacial deposits mapped within Sinks Canyon are most obvious as a series of lateral moraines along the ~6 km mid-portion of inner Sinks Canyon, from the Missouri Geology Camp to PopoAgie Falls (Fig. 4). In this reach of the canyon they stand out above and beyond the inner-most complex of Pinedale deposits on both sides of the canyon (Dahms, 2004a).

Our new ^{10}Be exposure ages for Bull Lake allostratigraphic units in Sinks Canyon (Dahms, 2004a) are generally similar to those ages reported previously (using older production rates and scaling models) by Chadwick et al. (1997) and Phillips et al. (1997) from the Bull Lake type Area (BLTA). Boulders N-1, N-2 and N-3 on the moraine/outwash complex outside the mouth of Sinks Canyon (Fig. 3; Table 1) yield exposure age-estimates between 93 ka and 163 ka. Additionally, we re-calculated the ages from the two valley-side bedrock transects of Fabel et al. (2004), the upper portions of which are associated with Bull Lake moraine units (Dahms, 2004a; Figs 3, 4; Table 1). Our re-calculated ^{10}Be exposure-ages of ca. 130–to-69 ka (Table 2) support Fabel et al's (2004) earlier interpretation that the bedrock between the Bull Lake and Pinedale map limits have been more/less continuously exposed to cosmogenic radiation since the retreat of Bull Lake- or possibly Early Wisconsin-age (MIS-4) ice from Sinks Canyon (Dahms, 2004a; Fabel et al., 2004; Hall and Shroba, 1993, 1995; Colman and Pierce, 1986).

4.4 Pinedale (Late Pleistocene - LGM)

^{10}Be exposure ages from two boulders on the terminal moraine at the mouth of North Fork Canyon (Pine Bar Ranch, Fig. 2) indicate that the Pinedale glacier here abandoned its terminus ca. 22.5-23.3 ka (Table 1). From these ages, we estimate that the Pinedale glacier in Sinks Canyon (at The Sinks; Fig. 4) reached its terminal position no later than ca. 22.5 ka. Our results are similar to the mean exposure ages (as recalculated by Shakun et al., 2015) reported for LGM terminal moraines from the Fremont Lake Type Area (FLTA; 23.2-22.5 ka), the Colorado Front Range (Middle Boulder Creek, 20.6$\pm$1.3 ka; Green Lake, 21.8$\pm$2.4 ka; Clear Creek, 20.6$\pm$0.5 ka; Pine Creek, 23.5$\pm$1.5 ka), the Wallowa Mountains (24.2$\pm$1.1 ka), the Uinta Mountains (E. Fork Smith's Fork and S. Fork Ashley Creek (20.4$\pm$2.4 ka, 22.5$\pm$1.8 ka), the Ruby Mountains (22.7$\pm$2.1ka), and the Sonora Junction moraines of the Sierra Nevadas (21.5$\pm$0.8-22.5$\pm$2.8 ka). We also note that Shakun and Carlson (2010) have determined that the average global maximum ice extent was ca. 22 ka, while Clark et al. (2009) place the duration of the LGM from 26.5 to 19.0 ka.

4.5 Lateglacial–to-Holocene

The period from ca. 19-11.5 ka is generally recognized as a period of global deglaciation (Shakun et al., 2015). Moraines attributed to this period in the western United States generally are termed as 'recessional LGM' deposits (Thackray, 2008) as the majority of exposure ages earlier than ~16 ka are reported from lateral and end moraines in the lower valleys of trunk glaciers (see Licciardi et al., 2004; Munroe et al., 2006; Laabs et al., 2009; Leonard et al., 2017). 'Lateglacial' boulder exposure ages associated with lower/outer cirque moraines from the western U.S. often are correlated to the Younger Dryas (12.9-11.7 ka of Alley, 2000, Alley et al., 1993; also see Osborn et al., 1995; Davis et al., 2009; Munroe and Laabs, 2017; Menounos and Reasoner, 1997; Osborn and Gerloff, 1997).

A pattern of cold climate activity following the Heinrich-1 event (16.8 ka, Hemming, 2004) prior to the Younger Dryas is already noted in regions of the U.S. west (Clark and Bartlein, 1995; Benson et al., 1997) and exposure ages corresponding to cirque glacier activity at this time are common from locales in the European Alps (Böhlert et al., 2011; Darnault et al., 2012; Ivy-Ochs, 2015; Palacios et al., 2017; Makos et al., 2018). We are, however, aware of no previous accounts from the western U.S. mountains where cirque moraines are explicitly attributed to glacial activity during the 'Oldest Dryas' (17.5-14.7 ka; INTIMATE event GS-2.1a of Rasmussen et al., 2014). Mean boulder exposure ages (Shakun et al., 2015) from the Junction Butte and Deckard flats (Yellowstone) moraines (15.5±0.7, 15.8±1.3 ka;) and from the Outer and Inner Jenny Lake (Tetons) moraines (15.9±0.9, 14.8±1.2 ka), however, suggest ice was active in these valleys at this time. Leonard et al. (2017) also report 17-14 ka exposure ages from boulders on cirque and upper-valley moraines in the Sangre de Cristo Mountains equivalent to the cirque/valley positions we report below. Thus, while no correlations have been made from the western U.S. specifically to 'Oldest Dryas' glacial activity, evidence begins to appear for this equivalence. In the following section we present evidence for glacier activity during both the Younger and Oldest Dryas periods from three cirques in the southern WRR.

4.5.1 Temple Lake Valley

We noted above that the Temple Lake Alloformation has previously been interpreted to correspond to a glacial advance in the WRR during the Younger Dryas (Hack,1943; Moss, 1949, 1951; Holmes and Moss, 1955; Birkeland et al., 1971; Miller and Birkeland, 1974; Dahms, 2002, et al., 2010). Using the available data originally reported by Marcott (2011), we recalculated eight exposure ages from boulders on the terminal moraine of the Temple Lake Type Locality here using CRONUS 2.3 and the same time-dependent scaling (Lm) of Lal/Stone (Lal, 1991; Stone, 2000). The recalculated ages (Table 1) range from 12.4±1.2 ka to 16.5±1.6 ka [unweighted average (minus young outlier) = 15.3±1.5 ka]. This range of ages suggests the Type Temple Lake moraine most likely was deposited during INTIMATE event GS-2.1a (17.5-14.7 ka; Rasmussen et al., 2014; Shakun and Carlson, 2010; Ivy-Ochs, 2015), rather than during event GS-1 (Younger Dryas, 12.9-11.7 ka; Rasmussen et al., 2014; Alley, 2000).

4.5.2 Stough Basin & Cirque of The Towers

Discrete moraine units were previously identified both in Stough Basin and Cirque of the Towers. through the use of relative age-characteristics. These units were correlated to four Lateglacial–Holocene glacial advances (Figs 5, 7; Dahms, 2002; Dahms et al., 2010). The oldest/outer cirque moraines ('Temple Lake') were correlated to the Younger Dryas (12.9-11.6 ka; Alley, 2000), while the immediately younger/inner moraines ('Alice Lake') were associated with the first Neoglacial advance of the Holocene (ca. 5-6 ka). The youngest two units were correlated to the Black Joe (1-2 ka) and Gannett Peak (LIA) alloformations. We present here new ^{10}Be exposure-ages from three sets of Temple Lake and Alice Lake-age moraines in Stough Basin and Cirque of the Towers that revise the previous correlations of these allostratigraphic units.

Marcott (2011) recently reported a series of ^{10}Be boulder exposure ages from the outer ('Temple Lake') and middle ('Alice Lake') moraines in Bigfoot Lake cirque (Fig. 5; Table 1). Our recalculations of Marcott's fourteen exposure ages (CRONUS 2.3) show the outer ('Temple Lake') moraine was deposited ca. 14.3±1.4 ka while the middle ('Alice Lake') moraine was deposited ca. 11.0±1.1 ka. These unweighted mean exposure ages are significantly older than Dahms' original interpretations for these two moraines (Dahms, 2002; Fig. 6 in Dahms et al., 2010). Marcott, however, recently detected (personal written communication, 2017) errors in his original data table (Marcott, 2011: Table B2), particularly in the data for the middle moraine (Alice Lake). Accounting for these errors, Marcott relates that the arithmetic mean of the corrected ages are 14.9 ka for the Bigfoot Lake outer moraine and 13.9 ka for the Bigfoot Lake inner moraine (pers. written communication, 2017; Table 1).

In order to corroborate Marcott's revised ages for the Temple Lake and Alice Lake allostratigraphic units in Stough Basin, we sampled additional boulders from moraines mapped as 'Temple Lake' and 'Alice Lake' in Helen Lake cirque, immediately southeast and adjacent to Bigfoot Lake (Fig. 6 in Dahms et al., 2010). We obtained unweighted average exposure ages, respectively, of 15.6±1.6 ka (n = 2) and 13.2±1.3 ka (n = 2) (Fig. 5; Table 1). These exposure ages largely agree with Marcott's recalculated ages (above) from adjacent Bigfoot Lake cirque. When combined, these age estimates indicate the moraines mapped as 'Temple Lake' and 'Alice Lake' in Stough Basin were deposited ca. 15.6-14.9 ka and ca. 13.9-13.2 ka, respectively.

The relative age techniques used to differentiate the moraines in Stough Basin were also used to distinguish the 'Temple Lake' from the 'Alice Lake' moraines in Cirque of the Towers (Fig. 4-F in Dahms et al., 2010). Our six ^{10}Be exposure-ages from these moraines in Cirque of the Towers closely correspond to the ages we report from Stough Basin.

The exposure ages of 11.2±1.1 and 12.3±1.3 ka from boulders CT-3 and CT-4 on the moraine mapped as 'Alice Lake' (Fig. 4-F in Dahms et al., 2010; Fig. 8) are slightly younger than those we report from the 'Alice Lake' moraines in Stough Basin (13.9-13.2 ka). Exposure ages from boulders CT-1 and CT-2 of 15.8±1.5 and 15.1±1.5 ka (ave. = 15.5 ka; Table 1) on the moraine below Warbonnet Peak mapped as 'Temple Lake' and the ages of boulders CT-5 and CT-6 of 15.1±1.5 and 16.1±1.6 (ave. = 15.6 ka) on the moraine enclosing Lonesome Lake (Fig. 8) indicate these moraines most likely were formed synchronously. The most likely scenario is that the 'Temple Lake' moraine below Warbonnet Peak is a right lateral moraine of the glacier that reached from the southern cirque headwall down to Lonesome Lake. It is possible, however, that the moraine enclosing Lonesome Lake was formed by ice flowing exclusively from the smaller cirques surrounding Pingora Peak while the southern moraine is a terminal moraine formed by a separate ice mass flowing from the Warbonnet Peak headwall (Fig. 8). The latter scenario would require the receding North Fork trunk glacier to have separated above Lonesome Lake into two discrete cirque glaciers prior to ~16.0 ka. The ice masses from the headwall above Pingora Peak would then have re-advanced to the distal end of Lonesome Lake while the ice mass from the southern cirque area advanced only far enough to deposit the southern 'Temple Lake' moraine. The mean exposure age (n = 4) of 15.5±1.5 ka from these two 'Temple Lake' moraines (Table 1; Figs 7, 8) correspond closely to the mean ages we report from the 'Temple Lake' moraines in Stough Basin [14.9 ka in Bigfoot cirque (Marcott, pers. written comm., 2017); 15.6 ka in Helen cirque](Figs 5, 8; Table 1).

The general similarity of exposure ages from moraines correlated to the Alice Lake and Temple Lake alloformations in Stough Basin and Cirque of the Towers indicates that these deposits were formed by two advances of cirque glaciers during the Lateglacial period (post-LGM/pre-Holocene). The similarity among the ^{10}Be exposure ages (Table 1; Fig. 9) from the 'Alice Lake' moraines in Stough Basin (13.9-13.2 ka) and in Cirque of the Towers (11.8 ka) indicate that the Alice Lake Alloformation in the southern Wind River Range most likely was formed during the extended cooling period(s) of GI-1c3,2,1 through GS-1[commonly termed the intra-Allerød cold period (IACP)-Younger Dryas of 13.9-11.7 ka; Rasmussen et al., 2014; Alley, 2000; Shakun and Carlson, 2010; Yu and Eicher, 2001] and not during the Holocene 'Neoglacial' as earlier proposed by Dahms (2002; Dahms et al., 2010). Additionally, the recalculated mean exposure age of 13.3±0.6 ka for the Titcomb Lakes moraine of the northern WRR (Gosse et al., 1995a, b; Shakun et al., 2015) indicates this unit should also correspond to the Alice Lake Alloformation (as presently revised) rather than to the Temple Lake Alloformation (Dahms et al., 2010).

We suggest that the means of the ages we report from the outer/older moraines in Stough Basin (n = 9; ~15 ka) and Cirque of the Towers (n = 4; ~15.6 ka) closely correspond to the mean exposure age of the boulders on the Type Temple Lake moraine (n = 8; ~15.3 ka) at the Temple Lake Type Locality as recalculated from Marcott (2011). We propose that the Temple Lake Alloformation in the southern Wind River Range should now correspond to the GS-2.1a event of 17.5-14.7 ka (commonly termed the Oldest Dryas; Rasmussen et al., 2014; Ivy-Ochs, 2015; Shakun and Carlson, 2010) rather than to the Younger Dryas (Dahms et al., 2010 and references therein).

4.6 Post-LGM Recession

4.6.1 Sinks Canyon

Five new ^{10}Be ages on moraine boulders (Table 1) and eleven re-calculated ^{10}Be and ^{26}Al exposure ages from polished/striated bedrock surfaces inside the highest-mapped Pinedale lateral moraines (Table 2) constrain the rate at which the Sinks Canyon glacier receded 7.2 km from its Pinedale terminal position at the Sinks. The arithmetic averages of boulder exposure ages from end moraines Pd2 and Pd3 indicate the Pinedale glacier in Sinks Canyon had receded to 1.1 km behind its estimated LGM position at The Sinks by 21.0 ka and to 2.6 km up-valley by ca. 19.4 ka (Table 1; Fig. 4).

Two side-valley bedrock transects (Fig. 4; Fig. 5a and b; Fig. 5 in Fabel et al., 2004) located 5.5 km and 7.2 km up-canyon from the Pinedale LGM position originally were used to demonstrate the systematics of glacial erosion in bedrock-floored alpine valleys and to estimate rate(s) of bedrock erosion. We re-calculated the ^{10}Be and ^{26}Al ages of Fabel et al. (2004) using updated scaling estimates in order to identify when the basin floor at these localities became ice-free (Fig. 5; Table 2). The ^{10}Be exposure ages from the polished/striated bedrock surfaces along Transect B indicate that the ice surface at the valley side-wall was near 2450 m at the LGM. By ~21.8 ka the ice surface here had lowered to ~2420 m (#97-47/48). By ~ 19.0 ka the ice surface had lowered to below ~2330 m (#97-85). The ^{10}Be exposure ages along Transect A-A' indicate the ice surface at the valley-side was ice-free down to at least 2640 m (#97-113) by ~17.0 ka. ^{26}Al exposure ages indicate that the ice surface had lowered to below 2560 m near the valley floor at the base of A-A' by ~15.2 ka (#97-118, 120).

By using the two exposure ages of 22.5±2.3 and 23.3±2.3 ka from boulders on the Pinedale terminal moraine at Pine Bar Ranch at the mouth of North Fork Canyon (Fig. 2; Table 1), we estimate that the corresponding Pinedale glacier in Sinks Canyon had probably begun to recede from its terminus by 22.5 ka. The average of the three boulder exposure ages from Sinks Canyon moraine Pd2 (Fig. 4) indicates the glacier receded 1.1 km from its terminus to moraine Pd2 by 21.0 ka±2.0 at a rate between 0.31 and 1.1 m/yr. It receded up-valley another 1.5 km to Pd3 by 19.4±1.8 ka at a rate of 1.1 to 1.2 m/yr. By the time the ice surface was below the base of Transect A-A' at 15.3±1.7 ka the glacier had receded another 4.6 km at a rate near 1.1 m/yr. We estimate the glacier's overall recession rate from its terminus at the Sinks at 22.5 ka to 7.2 km up-canyon at the base of transect A-A' at ~15.3±1.7 ka was from 0.7 to 1.3 m/yr.

4.6.2 North Fork Canyon

The two ^{10}Be exposure ages of 23.3±2.3 and 22.5±2.3 ka from boulders (PB-1, PB-2) on the terminal moraine at the mouth of North Fork Canyon at Pine Bar Ranch (Fig. 2; Table 1) constrain the age of the Pinedale maximum here to ca. 23.0±2.3 ka. We derived two additional exposure ages (Dickinson Park-1, Dickenson Park-2) from a small (~3 m-high) lateral moraine directly below Dickinson Park near the juncture of the Dickinson Park & North Fork trails (Fig. 2). Using 22.5 ka as the age for the onset of recession of the North Fork glacier from its Pinedale terminus, the younger Dickinson Park exposure age (DP-2) suggests the North Fork glacier had receded 13.5 km up-canyon by ca. 16.4±1.6 ka; this suggests a recession rate of between 1.8 and 3.0 m/yr. If the age estimate for boulder DP-2 is realistic, then this rate is ~2x the recession rate we estimated for Sinks Canyon (above). We assume that the age of ~26.9 ka for the second boulder is anomalous. This small moraine lies near river-terrace level at the foot of a set of lateral moraines, so boulder DP-1 may have originated from an older/higher moraine and lodged on the edge of this moraine when the glacier's terminus probably was located less than ca. 0.1 km downvalley.

Further up-canyon, the range of exposure ages of the three boulders above Lizard Head Meadows (Fig. 8) indicates that moraines from two separate glaciers are present at this locality. Our sampling area was located at the confluence where the Bear Lake cirque valley meets the North Fork valley (Fig. 8). Boulders CT-8 (14.8±1.5 ka) and CT-9 (14.9±1.5 ka) were located ~200 m downvalley from boulder CT-7 (17.6±1.7 ka). Boulders CT-8 and CT-9 thus appear to represent a marginal moraine deposited by the Bear Lake cirque glacier that contacted the left lateral moraine of the North Fork glacier after the North Fork glacier receded from this position. The ages of boulders CT-8 and CT-9, thus, suggest this moraine is of late Temple Lake-age. Boulder CT-7, therefore, represents a position of the North Fork trunk glacier as it receded up North Fork Canyon into the Cirque of the Towers. This recessional sequence corresponds well with the 16.1 – 15.1 ka ages reported from boulders CT-5 and CT-6 on the moraine enclosing Lonesome Lake. If boulder CT-7 represents a recessional position of the North Fork glacier at ~17.6 ka, then the

North Fork glacier receded 27 km up-canyon from its LGM position at Pine Bar Ranch at ~22.5 ka to this position in Lizard Head Meadows at a rate of ~5.0 m/yr (5000 m/kyr). This recession rate is 5x the rate we calculate for the Middle Fork (Sinks Canyon) glacier. In theory, valley glaciers with smaller ice-shed areas should respond more quickly to changes in climate conditions than those with larger areas (Davis et al., 2009). Thus, the higher recession rate of the North Fork trunk glacier could be due to either its smaller ice-shed area (102 km^2), when compared to that of the Middle Fork trunk glacier (149 km^2; Fig. 2) or differences in the hypsometry between the basins (Young et al., 2011), as the Cirque of the Towers-North Fork basin is generally deeper than Middle Fork basin.

4.6.3 Stough Basin

The Stough Basin valley trunk glacier at the LGM was composed of three coalescing cirque glaciers from the Ice Lake, Bigfoot Lake and Helen Lake cirques (Figs. 2, 6). The glacier from Helen Lake cirque occupied the eastern half of the basin; glaciers from Bigfoot Lake and Ice Lake cirques occupied the middle and western areas, respectively. A medial moraine located near the basin's center (Figs 6, 7a) apparently represents the location where the Helen Lake cirque glacier merged with the glacier from Bigfoot Lake cirque to form most of the ice volume in Stough Basin. Riegels at 3383 m and 3322 m impound the outer lakes in Helen and Bigfoot cirques, respectively.

With an area of 1.28 km^2, Helen cirque is the largest of the three cirques that fed ice to Stough Basin; Bigfoot and Ice cirques are progressively smaller (0.83 and 0.19 km^2, respectively). As a result, the Helen cirque ice stream most likely had the largest volume. We can estimate the upper limit of the ice stream from the position of a remnant lateral moraine on the eastern valley wall (see 97-90, 97-91 of Fig. 6). The upper elevation of this moraine remnant suggests that the surface of the ice stream was no higher than ca. 3340 m at this location. The absence of erratic boulders above the moraine suggests the Stough Basin ice surface may have reached no higher than ca. 3360 m during either Pinedale or Bull Lake glaciations.

Exposure ages from the two erratic boulders (16.0±1.7 and 15.9±1.6 ka) on the lateral moraine remnant (Fig. 6; Table 2), along with the 17.9±1.8 ka age of A-A' bedrock sample 97-68 (Fig. 7a) suggest the ice surface in the basin still maintained much of its LGM elevation/thickness at ~18.0

ka, the same period the Middle Fork trunk glacier was retreating 6+ km from its terminus in Sinks Canyon. The Stough Basin valley glacier apparently began to actively waste after ~18 ka (97-68). Consequently, ice decay between 20 ka and 18 ka (#97-90, #91) was extremely fast at lower elevations (cf. Rigual-Hernández et al., 2016; Monegato et al., 2017) which finally translated to high elevations after this time. [Note that lateral moraines generally form below ELA, so that the moraine with samples 97-90 and 97-91 was probably near or below ELA by ~16.0 ka.] Ages from the progressively lower bedrock surfaces along transects A and B (Fig. 7a, 97-66, 97-63; Fig. 7b, 97-97) indicate the surface elevation of the eastern ice stream emanating from Helen cirque decreased by ca. 70 m, from 3350 –to- 3280 m, between 17.9 ka and 13.5 ka (ca. 20 m to 3330 m by 14.8 ka and by another 50 m, to 3280 m, by 13.5 ka). Regardless of whether the true age of the Temple Lake-age moraine in Helen cirque is closer to the youngest boulder exposure age (13.9$\pm$1.5 ka) or to the oldest (~17.2$\pm$1.7 ka), the ages suggest that the ice immediately down-valley from the riegel (Fig. 7b, 97-97) lingered until sometime after ~14 ka. The ages of the boulders on the Temple Lake-age moraine in Helen cirque indicate the moraine was deposited by at least 13.9 ka. Thus, the ice in Helen cirque must have become detached across the riegels from the valley glacier below before 16 ka for ice to have retreated behind the Helen Lake riegel far enough to re-form the moraine. A similar situation is reported from the Uinta Mountains where downwasting ice is suggested to have exposed higher cirque areas while active ice remained in the lower valleys about this time (Refsnider et al., 2008).

On the west side of the basin the exposure age of 27.3$\pm$2.9 ka from the highest, western-most sample of transect A-A’ (Figs. 6, 7; 97-100) indicates the elevation of the Pinedale maximum ice surface in Ice Lake cirque was probably no higher than ca. 3460 m. The ~18.5 ka exposure age of transect B-B’ sample 97-75 indicates that as the ice surface here lowered following the LGM, ice from Bigfoot cirque still extended over its northern wall and remained connected to the Ice Lake ice stream. The mean of the bedrock exposure ages for 97-50, 97-52 and 97-55 (Fig. 7a) indicates that by ~18.1 ka the Ice Lake ice surface had lowered below 3380 m. By ca. 17.0-16.5 ka the surface of the ice flowing from Bigfoot cirque into Wilhelm Lake had lowered enough (from 3370 m to 3300 m) to expose the ridge separating Wilhelm Lake and Ice Lake (Fig. 6; Fig. 7b, 97-76; Fig. 7a, 97-57, 59).

The exposure ages from samples 97-78 and 97-80 (Fig. 7b) suggest that ice had retreated from the Bigfoot cirque riegel (~3330-3290 m) by at least 16.1 ka and possibly as early as 17.6 ka. Thus, while the Helen Lake cirque glacier remained coupled to its down-valley ice stream until ~16-15 ka, ice in Bigfoot Lake cirque had already detached across the Bigfoot cirque riegel from the remaining ice in the lower valley by 17-16 ka. Near the basin's center, the exposure age from 97-62 (Figs. 6, 7a) indicates that stagnant ice remained on the floor of the basin east of Wilhelm Lake until 12.5$\pm$1.3 ka.

The decoupling of ice at the Bigfoot cirque riegel by 17-16 ka suggests that the ELA was not much higher than ~3320 m (10,900') at this time -- a level that would allow relatively large areas of the main valley glacier system (Deep Lakes, Ice Lakes) to remain above this elevation. It appears that the surface of the Stough Basin ice stream quickly lowered and stagnated following the LGM while the Middle Fork trunk glacier remained active with ice from the Deep and Ice lakes systems, although in negative mass balance. This relationship supports the observation that the Stough Basin glacier was no more than a secondary source of ice to the Middle Fork-Sinks Canyon trunk glacier (and probably stopped contributing ice to the Middle Popo Agie trunk glacier about this time), while the main accumulation areas were located in the Deep Creek Lakes/Ice Lakes cirque valleys to the northwest (Fig. 2).

The Bigfoot and Helen cirque riegels are ca. 18.8 km upvalley from the Pinedale glacier's LGM terminus at the Sinks. Our exposure ages indicate that the terminus of the PopoAgie trunk glacier still extended to near transect A-A' (Fig. 5a) at ~15.3 ka in Sinks Canyon, while the ice surface in Stough Basin had already lowered enough to expose the Bigfoot cirque riegel by ~17 ka. Thus, while the Middle Fork glacier in Sinks Canyon receded relatively slowly from its terminus to transect A-A' (0.7-1.3 m/yr), the overall rate of down-wasting, as calculated from the estimated age when ice began to recede from the Pinedale terminus at 22.5 ka to the time ice abandoned the Helen and Bigfoot cirque riegels ($\geq$16 ka) is 2.9 m/yr.

5. Conclusions

5.1 Sinks Canyon – Table Mountain

Our ^{10}Be and ^{26}Al exposure ages from moraine boulders and polished/striated bedrock from the Middle Fork and North Fork basins of the PopoAgie River more tightly constrain the sequence of Pleistocene glacial activity previously reported for the PopoAgie Basin (Dahms, 2002, 2004a; Dahms et al., 2010). Exposure ages from moraine boulders on alloformations in/near Sinks Canyon correspond to at least four periods of glacial activity during the Early, Middle, and Late Pleistocene. The oldest boulder exposure ages of 1.0-2.0 Ma from the Table Mountain diamicton previously identified as an undated Older pre-Sacagawea Ridge deposit (Dahms, 2004a) now likely relates this diamict to an episode of Early Pleistocene pre-canyon glacial activity. We propose the Table Mountain diamicton should be identified as the Table Mountain Alloformation and associated with at least one Early Pleistocene pre-canyon glaciation.

The single boulder exposure age of ca. 556 ka, when combined with the geomorphic characteristics of the corresponding patches of isolated boulders and diamicts extending from below Table Mountain onto and along the south rim of Sinks Canyon (Figs 2, 3) tentatively suggests that Dahms' (2004a) original 'Younger pre-Sacagawea Ridge' unit now corresponds more closely to the Sacagawea Ridge Alloformation. When combined with previous studies from the eastern WRR that describe outwash terraces containing Lava Creek B tephra (~650 ka) that grade to the Sacagawea Ridge moraines at Dinwoody Lakes and near the BLTA (Richmond and Murphy, 1965; Richmond, 1976; Chadwick et al., 1997), our boulder exposure age and new Lava Creek ash locality at the Lander airport terrace (Dahms and Egli, 2016; William McIntosh, personal written communication, January 2018) indicate this diamict more than likely corresponds to the Sacagawea Ridge Alloformation and is most likely of MIS-16 age (Cohen and Gibbard, 2011). However, in light of its large external uncertainty, taken by itself our ^{10}Be exposure age could represent a span of MIS stages from 10 to 18.

Exposure ages from boulders on the moraines at the mouth of North Fork Canyon (Fig. 2) indicate the Pinedale valley glacier advanced to the present location of the Pine Bar Ranch by ca. 23 ka. This age closely corresponds to the recalculated ages (Shakun et al., 2015) of ~20-23 ka from the oldest distal LGM moraines at the Fremont Lake Type Area (Wyoming), the Colorado Front Range, the Sierra Nevadas and ages from the Uinta Mountains (Munroe and Laabs, 2017) -- all of which generally correspond to Shakun and Carlson's (2010) estimate of 22-23 ka for the global LGM.

Boulder exposure ages from moraines outside and within Sinks Canyon (Table 1; Fig. 4) and re-calculated ages from polished-striated bedrock along two valley-side transects (Fig. 4; Table 2; Fabel et al., 2004) corroborate their previous associations with the Bull Lake and Pinedale alloformations (Dahms, 2004a) and with MIS-6/5 and MIS-2, respectively. We were unable to find datable boulders from moraines previously mapped by Dahms (2004a) as Early Wisconsin (MIS4) and so the presence here and the age of this alloformation remains uncertain.

5.2 Stough Basin and Cirque of the Towers

Relations among the exposure ages from the two boulders on the lateral moraines on the east wall of Stough Basin (Tables 1 and 2; Fig. 6), the bedrock ridge between Ice and Wilhelm lakes and the moraines in Helen and Bigfoot cirques suggest the ice streams emanating from them began to stagnate after ~18 ka. Stagnant ice probably remained on the southern floor of Stough Basin until ~14-12 ka, but active cirque ice had decoupled across the cirque riegels and retreated behind the cirque riegels by 17-16 ka. In a similar manner, exposure ages on the moraine boulders in Lizard Head Meadows suggest that the North Fork trunk glacier retreated far enough into Cirque of the Towers after 17 ka for it to have re-advanced to form the Lonesome Lake moraines by ~15.6 ka.

Boulder exposure ages from Stough Basin and Cirque of the Towers (Table 1; Fig. 9) show that the outer-most cirque moraines in the southern WRR were formed during the Lateglacial Pleistocene rather than the Holocene. The exposure ages from Bigfoot, Helen, Cirque of the Towers and the Temple Lake Type Locality indicate that the Temple Lake Alloformation in the WRR now should be considered coeval with the INTIMATE GS-2.1a cooling event of 17.4-14.7 ka commonly known as the Older Dryas (Rasmussen et a., 2014; Benson et al., 1997; Shakun and Carlson, 2010; Ivy-Ochs, 2015) rather than with the Younger Dryas, as earlier interpreted by Dahms (2002; Dahms et al., 2010 and references therein). Following a recession, presumably during the Bølling-Allerod interstadial, these cirque glaciers re-advanced in response to the IACP-Younger Dryas stadial(s) to form moraines of the Alice Lake Alloformation. Thus, we revise Dahms' earlier correlation of the Alice Lake alloformation with the mid-Holocene 'Neoglacial' period (2002; et al., 2010) and propose that the deposits corresponding to the Alice Lake Alloformation in Stough Basin and Cirque of the Towers are most likely coeval with the extended IACP-Younger Dryas cooling episode (Fig. 9) and it is possible that evidence no longer exists for early- to mid-Holocene glacial events in the southern Wind River Range.

The above suggested equivalences to the Oldest Dryas and Younger Dryas Lateglacial stadial events are supported by the $\delta^{18}O$ data derived from the GISP2 and NGRIP ice cores (Stuiver et al., 1995; Grootes and Stuiver, 1997; Rasmussen et al., 2014), the Owens Lake sediment cores (Benson et al., 1997). Cirque moraines reported to result from glacial activity during the Younger Dryas are common from the western U.S. (Osborn et al., 1995; Davis et al., 2009). Explicit reports of cirque moraines attributed to activity during the Older Dryas period are few in the western U.S. (although recalculations of previous exposure ages suggest equivalents may exist) while reports of moraines corresponding both to the Egesen (Younger Dryas equivalent) and Gschnitz (Older Dryas equivalent) advances of the European Alps are common (e.g., Darnault et al., 2012; Böhlert et al., 2011; Ivy-Ochs, 2015; Palacios et al., 2017; Makos et al., 2018).

In order to refine and test the ideas presented here and place them within a coherent record of landscape evolution, future work in this region should focus on:

(1) Identification of new exposures of early Pleistocene deposits. The geomorphic expression of the Table Mountain diamicton suggests that more than one allostratigraphic unit may be represented here. Additional paleomagnetic and numeric age-analyses should be applied to isolated high-elevation landforms in situations similar to Table Mountain outside the canyon mouths (e.g., Veggian et al., 2010);
(2) More accurate exposure ages for the Sacagawea Ridge allostratigraphic unit. The presently available evidence makes it possible to place this unit anywhere from MIS-10 to MIS-18;
(3) Additional exposure ages from moraines and bedrock in the Deep Lakes and Ice Lakes basins in order to estimate the recession rate of the Middle Popo Agie trunk glacier;
(4) Additional western U.S. localities for the rate(s) of recession of valley glaciers from their terminal LGM position(s).
(5) Additional exposure age-estimates from the Type Alice Lake moraine of the southern Wind River Range;
(6) Additional well-dated evidence of the geomorphic and stratigraphic record of all post-LGM advances in the region, how well these records represent a global record of deposits formed in response to Lateglacial stadial and inter-stadial climate cycles (e.g., Heinrich

events; the Blytt-Sernander sequence), the Holocene record of climate cycling (the 8.2 ka event; the 'Neoglacial') and regional-to-global rhythms of climate change (e.g., Clark and Bartlein, 1995; Yu and Eicher, 1998, 2001; Shakun and Carlson, 2010; Ivy-Ochs, 2015).

(7) ELA reconstructions and glacial modelling, in order to further test some of our proposed correlations and deglaciation geometries.

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Figure Captions

Fig. 1. Locations of the study area in North America and the Middle/ North Fork basins of the PopoAgie River system in the southeastern Wind River Range, including Stough Basin, Sinks Canyon and Cirque of the Towers. Detailed image from Landsat TM (Path 36, Row 30 (rgb),

2% linear, 250k).

Fig. 2. Locations of the main upper-basin areas contributing ice to the trunk glaciers of the Middle and North Fork basins of the Popo Agie River system at the southeastern flank of the Wind River Range. The lower limit of each area approximates the confluence of the lowest cirque-valley ice stream with the canyon's trunk glacier. The ice-shed for the Middle Popo Agie system is ca. 149 km^2 while the ice-shed for the North Fork system is ca. 102 km^2. Note the locations of the Temple Lake and Alice Lake type localities. (Google Earth image).

Fig. 3. Oblique view of Table Mountain in relation to the mouth of Sinks Canyon and the allostratigraphic units of the Middle PopoAgie Basin noted in this study. Orange = Table Mountain diamicton; Solid yellow = units of Sacagawea Ridge till; Dotted yellow = isolated Sacagawea Ridge stagnant ice deposits within mapped Bull Lake boundary; Single yellow dots = isolated erratic boulders correlated to Sacagawea Ridge; Solid red = Bull Lake moraine and stagnant ice; Dashed red = assumed past limit of Bull Lake lateral moraine; Green solid/dashed = Pinedale moraine limit/purported past position. Red dots = location of boulders sampled for ^{10}Be age analysis. HR1 & HR2 = locations of soil profiles noted in Dahms (2004a). Dashed orange lines = suggested kame terrace remnants on the NE of Table Mountain. (Google Earth image).

Sampled boulders: TM1-6 = Table Mountain; ET1-2 = East Table Mountain; N1–3 = Nicholas Ranch; DS1 = Deer Spring. (Google Earth image).

Fig. 4. Vertical view of Sinks Canyon showing relations among previously mapped Sacagawea Ridge (yellow), Bull Lake (red) and Pinedale (green) moraine units (Dahms, 2004a) and isolated boulders sampled for ^{10}Be exposure ages (yellow dots). Valley-side transects A and B indicated by white dots as locations of bedrock/boulder samples for ^{10}Be and ^{26}Al exposure ages. (Google Earth image).

Fig. 5. (a) Details of Middle PopoAgie valley-side transects A-A' and (b) B-B' showing sample locations, sample numbers, and recalculated exposure ages (Table 2). All exposure ages are from polished-striated bedrock outcrops except triangles that designate samples taken from boulders (samples A-A', 97-110 and 97-117. Asterisk above age ranges designates that exposure ages are given as ^{10}Be, ^{26}Al ages. Figure base modified from Figure 5 of Fabel et al. (2004).

Fig. 6. Oblique view of Stough Basin showing locations of Alice Lake (blue) and Temple Lake (red)

moraines in Helen Lake and Bigfoot Lake cirques and associated locations of dated boulders. Cross-valley transects A-A' ad B-B' are highlighted in orange. White dots indicate locations of bedrock samples for ^{10}Be and ^{26}Al exposure ages. Dotted white lines locate the mid-valley medial moraine (east) separating the Bigfoot Lake ice stream from the Helen Lake ice stream and the bedrock ridge (west) separating the Ice Lake from the Bigfoot Lake ice streams. Locations 97-90 and 97-91 are ^{10}Be-dated boulders on a remnant lateral moraine. (Google Earth image).

Fig. 7. Details of Stough Basin cross-valley transect A-A' (a) and B-B' (b) showing sample locations, sample numbers, and recalculated apparent exposure ages sampled from polished-striated bedrock exposures (Table 2).

Fig. 8. Vertical view of Cirque of the Towers showing major features of the cirque and the moraine units sampled for this study. Red = Temple Lake moraines; Blue = Alice Lake moraines; Yellow = medial moraine at the head of Lizard Head Meadows, between the glacier emanating from Cirque of the Towers and the glacier from Bear Lake. Black and white dots = locations of sampled boulders with associated sample numbers (Table 1). Note that the Alice Lake moraine south of Pingora Peak was not sampled but is included as the only other 'Alice Lake' moraine unit mapped here (Dahms et al., 2010). (Google Earth image).

Fig. 9. Summary of chronologies for Lateglacial-Holocene glacial activity in the Wind River Range [Fremont Lake Type Area (FLTA), Sinks Canyon, North Fork Canyon, Helen Lake, Bigfoot Lake, Temple Lake Type Locality (Marcott, 2011) and Titcomb Basin] normalized to the GISP2 2-m δ^{18}O record (Stuiver et al., 1995). Error bars approximate 1σ uncertainties. Grey dots represent individual ^{10}Be exposure ages from Alice Lake moraines; black dots represent individual exposure ages from Temple Lake moraines. Black triangles represent exposure ages from terminal moraines at the FLTA and Pine Bar Ranch; open triangles represent ages from moraines PD-2 and PD-3 in Sinks Canyon and moraines DP-1 and DP-2 in North Fork Canyon. Solid grey and black squares represent unweighted arithmetic averages of ages on moraine groups mapped as 'Alice Lake' and 'Temple Lake'. YD = Younger Dryas (GS-1 event of Rasmussen,et al., 2014); IACP = Intra-Allerød Cold Period; BA = Bølling/Allerød; OD = Oldest Dryas (GS-2.1a event of Rasmussen et al., 2014).

Ages for the Fremont Lake Type Area (FLTA) and Titcomb Lake are from Shakun et al. (2015). Average ages for the SE Alps are generalized from Ivy-Ochs (2015; personal oral communication, 2016) and Böhlert et al. (2011). The age of Heinrich Event-1 is from Hemming (2004) and Rood et al. (2011).

Table 1
Exposure ages of moraine boulders, Middle and North Forks PopoAgie Basin, Wyoming.

Unit (*specification*)	Sample #	Latitude (°N)	Longitude (°W)	Elevation (m a.s.l.)	Sample thickness (cm)	Shielding factor	Erosion rate (mm ky^{-1})	^{10}Be Exposure Age (ka)[1)]
Table Mountain	Table Mtn-1	42.76	-108.76	2219	2.0	1.000	2.00	1397.3 ± 6903.6
	Table Mtn-2	42.76	-108.76	2219	2.0	1.000	2.00	149.1 ± 22.7 (14.5)
	Table Mtn-3	42.76	-108.76	2225	2.0	1.000	2.00	404.2 ± 97.2 (51.4)
	Table Mtn-4	42.76	-108.76	2227	2.0	1.000	2.00	199.1 ± 35.9 (26.2)
	Table Mtn-5	42.74	-108.77	2243	2.0	1.000	2.00	373.6 ± 95.1 (72.7)
	Table Mtn-6	42.74	-108.77	2253	2.0	1.000	2.00	Saturated
	East Table-1	42.76	-108.73	2225	4.0	0.993	2.00	429.6 ± 106 (44.7)
	East Table-2	42.76	-108.74	2225	3.0	1.000	2.00	235.3 ± 38 (16.4)
Sacagawea Ridge	Deer Spring	42.73	-108.81	2300	2.5	0.993	2.00	556.4 ± 187.7 (103.9)
Bull Lake	Nicholas-1	42.46	-108.47	1783	2.0	0.921	2.00	126.6 ± 15.9 (5.3)
Outer Sinks Canyon	Nicholas-2	42.46	-108.47	1780	3.0	0.999	2.00	163.3 ± 22.1 (7.5)
	Nicholas-3	42.46	-108.47	1777	3.0	0.963	2.00	92.6 ± 11.8 (6.1)
Pinedale	Pinedale 2-1	42.74	-108.83	2083	2.5	0.980	0.00	19.7 ± 1.9 (0.5)
Lower Sinks Canyon	Pinedale 2-2	42.74	-108.83	2086	3.0	0.983	0.00	21.4 ± 2.1 (0.7)
	Pinedale 2-3	42.74	-108.83	2086	2.5	0.983	0.00	21.8 ± 2.1 (0.6)
	Pinedale 3-1	42.74	-108.85	2168	1.5	0.973	0.00	18.7 ± 1.8 (0.3)
	Pinedale 3-2	42.74	-108.85	2174	2.5	0.977	0.00	20.1 ± 1.9 (0.5)
North Fork Canyon	Pine Bar-1	42.87	-108.90	1901	3.5	0.885	0.00	23.3 ± 2.3 (0.9)
	Pine Bar-2	42.87	-108.90	1901	2.4	0.959	0.00	22.5 ± 2.3 (0.9)
	Dickinson Pk1	42.81	-109.05	2623	2.8	0.994	0.00	26.9 ± 2.7 (0.9)

	Dickinson Pk2	42.81	-109.05	2624	0.6	0.994	0.00	16.4 ± 1.6 (0.6)
	CT7	42.78	-109.19	3086	3.2	0.943	0.00	17.6 ± 1.7 (0.5)
	CT8	42.78	-109.19	3086	2.0	0.950	0.00	14.8 ± 1.5 (0.6)
	CT9	42.78	-109.18	3062	2.0	0.982	0.00	14.9 ± 1.5 (0.5)
Temple Lake	Helen TL-1	42.64	-109.01	3399	3.0	0.962	0.00	13.9 ± 1.5 (0.7)
Helen Lake cirque	Helen TL-2	42.64	-109.01	3354	3.0	0.962	0.00	17.2 ± 1.7 (0.6)
Cirque of the Towers	CT1	42.77	-109.22	3216	2.5	0.947	0.00	15.8 ± 1.5 (0.5)
	CT2	42.77	-109.22	3214	3.2	0.954	0.00	15.1 ± 1.5 (0.5)
	CT5	42.73	-109.21	3125	3.0	0.972	0.00	15.1 ± 1.5 (0.5)
	CT6	42.73	-109.21	3125	2.2	0.972	0.00	16.1 ± 1.6 (0.5)
Bigfoot Lake cirque	[M]SCO - 001	42.64	-109.01	3347	2.0	0.946	0.00	12.8 ± 1.3 (0.6)
	[M]SCO - 002	42.64	-109.02	3355	2.0	0.946	0.00	14.9 ± 1.4 (0.4)
	[M]SCO - 003	42.64	-109.02	3355	2.0	0.946	0.00	15.6 ± 1.5 (0.5)
	[M]SCO - 004	42.64	-109.02	3355	2.0	0.943	0.00	14.5 ± 1.6 (0.8)
	[M]SCO - 005	42.64	-109.02	3355	2.0	0.946	0.00	15.3 ± 1.6 (0.7)
	[M]SCO - 006	42.64	-109.02	3355	2.0	0.946	0.00	13.9 ± 1.4 (0.4)
	[M]SCO - 007	42.64	-109.02	3363	2.0	0.946	0.00	13.1 ± 1.3 (0.5)
Temple Lake Valley	[M]TLO-01	42.72	-109.18	3253	2.0	0.981	0.00	15.7 ± 1.5 (0.4)
	[M]TLO-02	42.72	-109.18	3253	2.0	0.987	0.00	16.5 ± 1.6 (0.5)
	[M]TLO-03	42.72	-109.18	3253	2.0	0.985	0.00	15.8 ± 1.6 (0.5)
	[M]TLO-04	42.72	-109.18	3253	2.0	0.988	0.00	15.5 ± 1.5 (0.4)
	[M]TLO-05	42.72	-109.18	3253	2.0	0.988	0.00	14.8 ± 1.5 (0.6)
	[M]TLO-06	42.72	-109.18	3253	2.0	0.984	0.00	12.4 ± 1.2 (0.5)

	[M]TLO-07	42.72	-109.18	3253	2.0	0.987	0.00	14.3 ± 1.5 (0.7)
	[M]TLO-08	42.72	-109.18	3253	2.0	0.987	0.00	14.8 ± 1.6 (0.8)
Alice Lake	Helen Al-1	42.63	-109.01	3399	1.5	0.957	0.00	13.6 ± 1.5 (0.7)
Helen Lake cirque	Helen Al-3	42.63	-109.01	3398	2.0	0.944	0.00	12.7 ± 1.2 (0.3)
Cirque of the Towers	CT3	42.77	-109.22	3225	1.8	0.930	0.00	11.2 ± 1.1 (0.4)
	CT4	42.77	-109.22	3225	3.0	0.935	0.00	12.3 ± 1.3 (0.5)
Bigfoot Lake cirque	[M]SCM - 01	42.64	-109.02	3370	2.0	0.937	0.00	12.2 ± 1.2 (0.3)
	[M]SCM - 02	42.64	-109.02	3370	2.0	0.937	0.00	10.7 ± 1 (0.3)
	[M]SCM - 03	42.64	-109.02	3370	2.0	0.935	0.00	9.6 ± 0.9 (0.3)
	[M]SCM - 04	42.64	-109.02	3370	2.0	0.937	0.00	12.1 ± 1.2 (0.5)
	[M]SCM - 05	42.64	-109.02	3370	2.0	0.937	0.00	12.2 ± 1.2 (0.3)
	[M]SCM - 06	42.64	-109.02	3370	2.0	0.936	0.00	9.9 ± 1 (0.3)
	[M]SCM - 07	42.64	-109.02	3370	2.0	0.937	0.00	10.2 ± 1 (0.4)

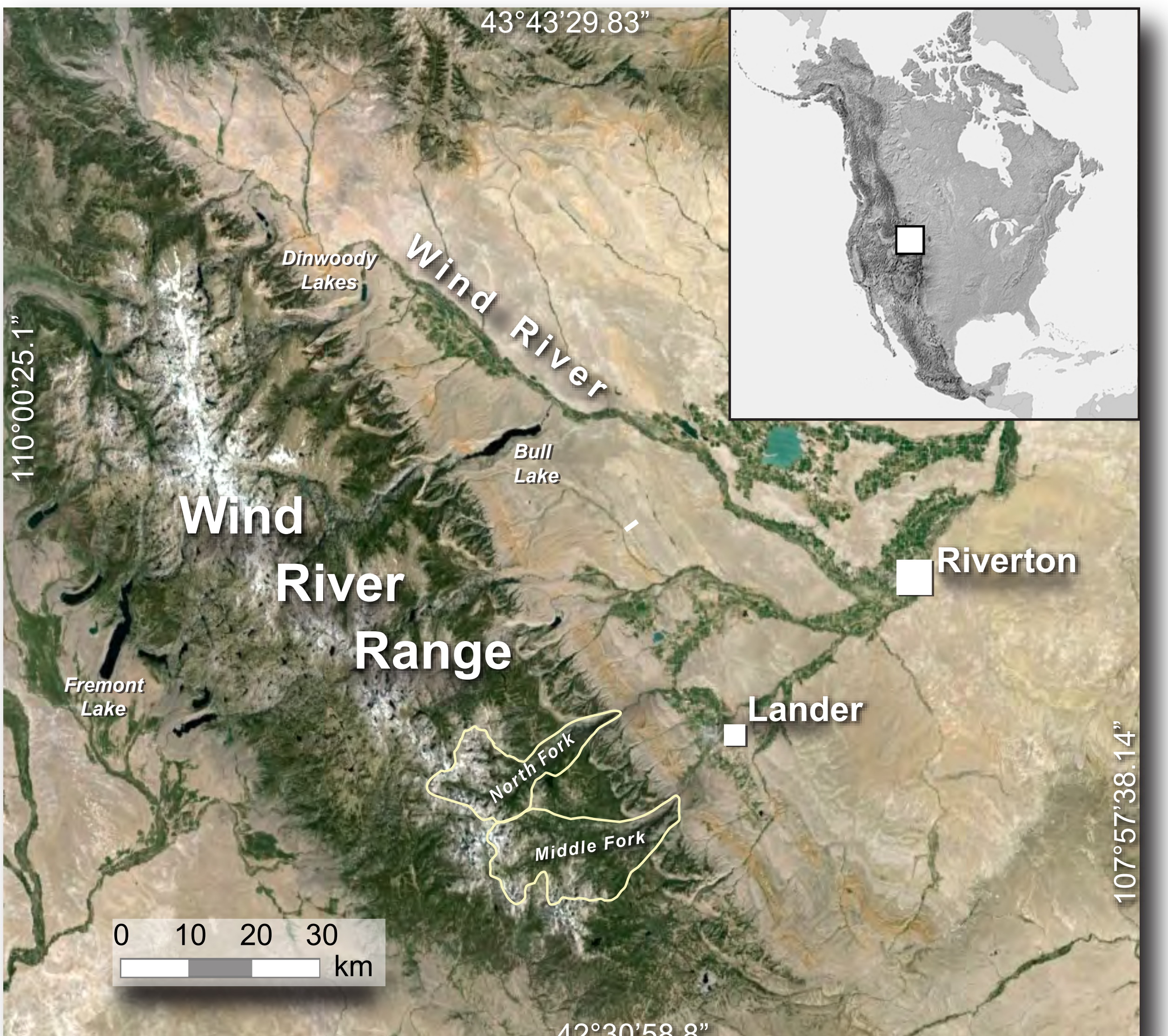
[M]data from Marcott (2011)
[1)] age with 1δ external and internal uncertainty

Table 2
Exposure ages (^{10}Be and ^{26}Al) for bedrock transects and boulders, Middle Fork PopoAgie Basin.

Unit	Sample #	Latitude (°N)	Longitude (°W)	Elevation (m a.s.l.)	Sample thickness (cm)	Shielding factor	Erosion rate (mm ky^{-1})	^{10}Be Exposure Age (ka) [1)]	^{26}Al Exposure Age (ka) [1)]
Upper Sinks Canyon	A-97-108^	42.73	-108.90	2830	1	0.492	0.0	130.9 ± 13.7 (5.8)	114.5 ± 13.7 (8.4)
	A-97-109	42.73	-108.90	2780	2	0.982	0.0	68.9 ± 6.9 (2.4)	54.8 ± 6.2 (3.6)
	A-97-110b^	42.73	-108.90	2720	2	0.997	0.0	17.2 ± 2 (1.2)	16.2 ± 1.9 (1.2)
	A-97-111^	42.73	-108.90	2680	1	0.997	0.0	21.2 ± 2.3 (1.1)	17.4 ± 2.3 (1.6)
Transect A	A-97-113^	42.73	-108.90	2650	1	0.979	0.0	17.5 ± 2 (1.2)	17.7 ± 2.1 (1.3)
	A-97-117b^	42.72	-108.90	2590	1	0.998	0.0		17 ± 2.1 (1.4)
	A-97-118^	42.72	-108.90	2560	1	0.965	0.0		15.1 ± 1.8 (1.1)
	A-97-120^	42.72	-108.90	2520	1	0.989	0.0		15.4 ± 1.7 (1)
	B-97-42^	42.73	-108.88	2560	2	0.998	0.0		63.2 ± 7 (3.7)
	B-97-43^	42.73	-108.88	2530	1	0.525	0.0	123.1 ± 12.4 (4.3)	
	B-97-44^	42.73	-108.88	2500	5	0.972	0.0	92.4 ± 9 (2.2)	
	B-97-45^	42.73	-108.88	2480	1	0.991	0.0	98.7 ± 10.4 (4.8)	98.3 ± 10.9 (5.6)
	B-97-46^	42.72	-108.88	2460	1	0.991	0.0	101.8 ± 10.9 (5.4)	
Transect B	B-97-47^	42.72	-108.88	2440	1	0.993	0.0	21.6 ± 2.5 (1.5)	
	B-97-48^	42.73	-108.88	2420	3	0.980	0.0	21.8 ± 2.6 (1.6)	
	B-97-49^	42.72	-108.87	2410	1	0.989	0.0	19.8 ± 2.5 (1.6)	
	B-97-85^	42.72	-108.88	2350	1	0.991	0.0	19.2 ± 2.6 (1.9)	18.4 ± 2.1 (1.3)
	B-97-88^	42.72	-108.88	2450	2	0.986	0.0	19.6 ± 2.2 (1.2)	
Stough Creek Basin	97-100	42.65	-109.02	3460	2	0.985	0.0		27.3 ± 2.9 (1.5)
	97-50	42.65	-109.19	3420	2	0.969	0.0	17.6 ± 1.8 (0.7)	
	97-52	42.65	-109.18	3390	2	0.970	0.0	18.4 ± 2 (1)	
	97-55	42.65	-109.01	3380	2	0.997	0.0	18.3 ± 1.8 (0.7)	
Transect A	97-57	42.65	-109.15	3340	2	0.997	0.0	16.5 ± 1.7 (0.6)	
	97-59	42.65	-109.11	3300	2	0.983	0.0	17 ± 1.7 (0.7)	
	97-62	42.65	-109.00	3250	2	1.000	0.0	12.5 ± 1.3 (0.6)	
	97-63	42.65	-109.00	3280	2	1.005	0.0	13.5 ± 1.4 (0.6)	

	97-66	42.65	-109.00	3330	2	1.011	0.0	14.8 ± 1.4 (0.4)	
	97-68	42.65	-108.98	3350	2	1.016	0.0	17.9 ± 1.8 (0.6)	
	97-75	42.64	-109.01	3390	2	1.022	0.0	18.5 ± 1.8 (0.6)	
	97-76	42.64	-109.01	3370	2	1.028	0.0	16.9 ± 2.1 (1.4)	
Transect B	97-78	42.64	-109.01	3330	2	1.033	0.0	16.1 ± 1.6 (0.6)	
	97-80	42.64	-109.00	3290	2	1.038	0.0	17.6 ± 1.8 (0.7)	
	97-97	42.65	-109.00	3320	2	1.044	0.0	14.7 ± 1.4 (0.4)	
Stough Lateral moraine	97-90b	42.66	-109.00	3320	2	1.050	0.0		15.9 ± 1.6 (0.7)
	97-91b	42.65	-109.00	3330	2	1.055	0.0		16 ± 1.7 (0.8)

^ Original data published in Fabel et al. (2004)
b boulder
1) age with external and internal uncertainty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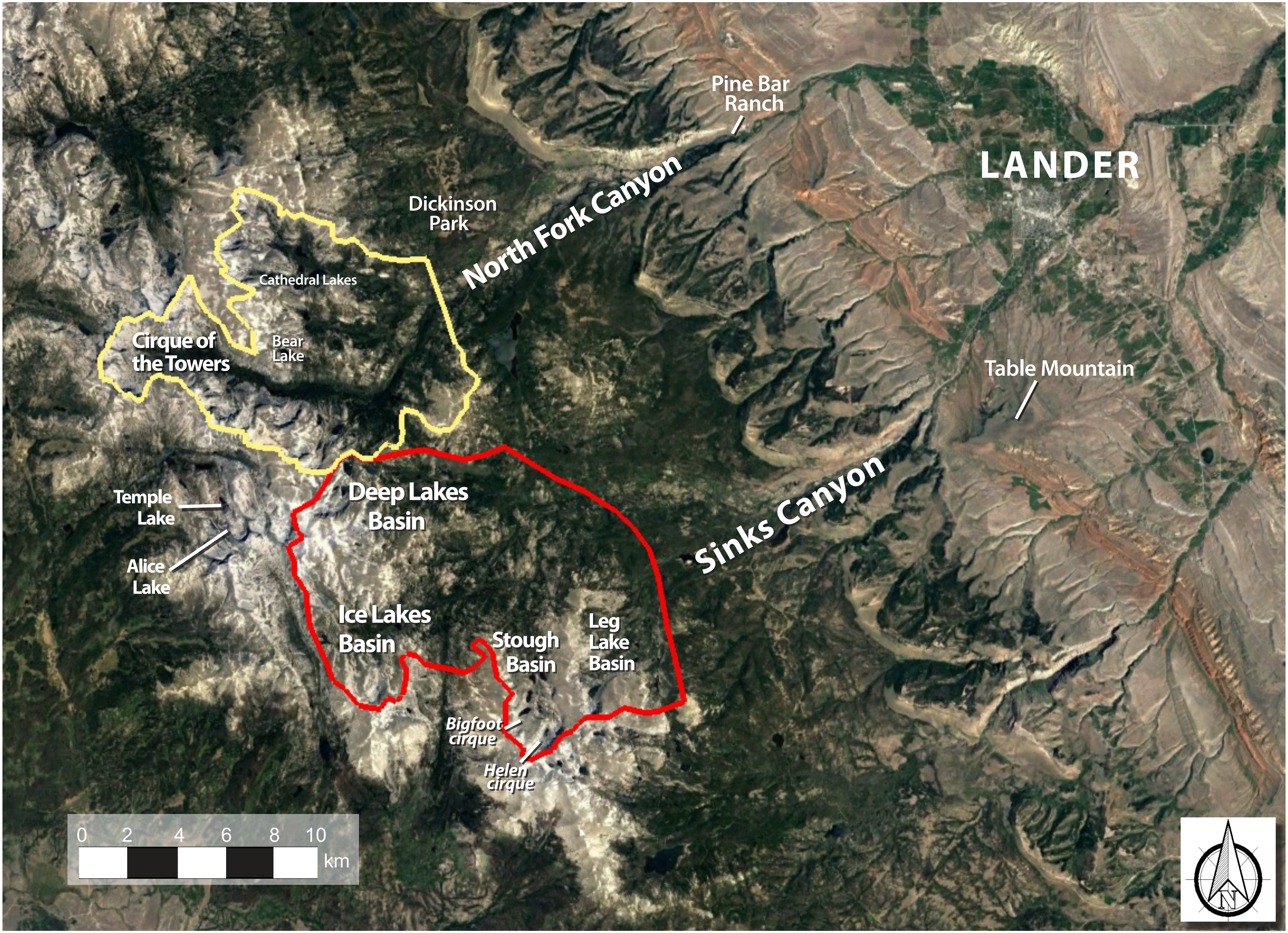
Wind River
Dinwoody Lakes
Bull Lake
Wind River Range
Riverton
Lander
Fremont Lake
North Fork
Middle Fork
0 10 20 30
km


Pine Bar
Ranch
LANDER
North Fork Canyon
Dickinson
Park
Cathedral Lakes
Cirque of
the Towers
Bear
Lake
Table Mountain
Deep Lakes
Basin
Sinks Canyon
Temple
Lake
Alice
Lake
Ice Lakes
Basin
Stough
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Leg
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Bigfoot
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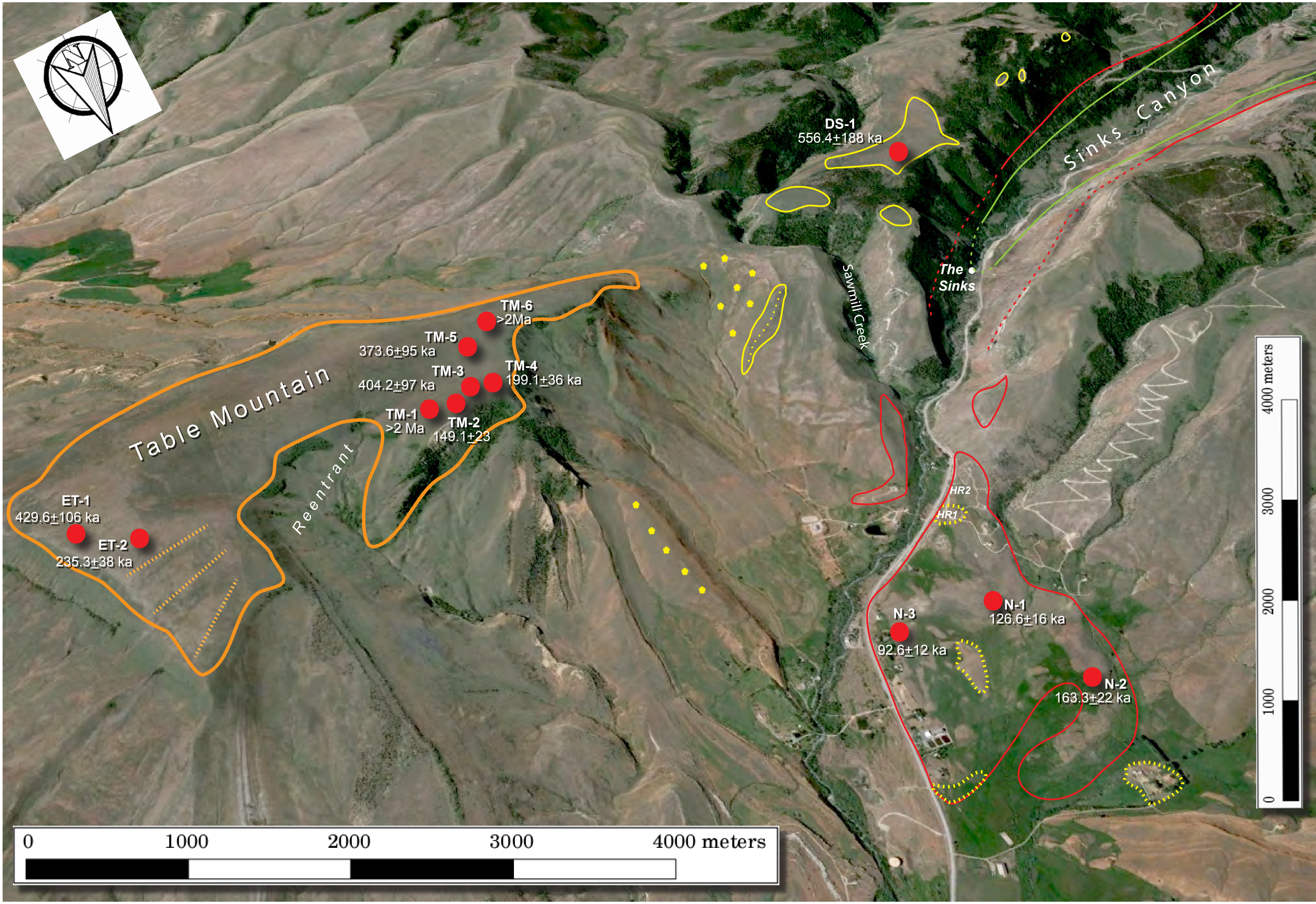
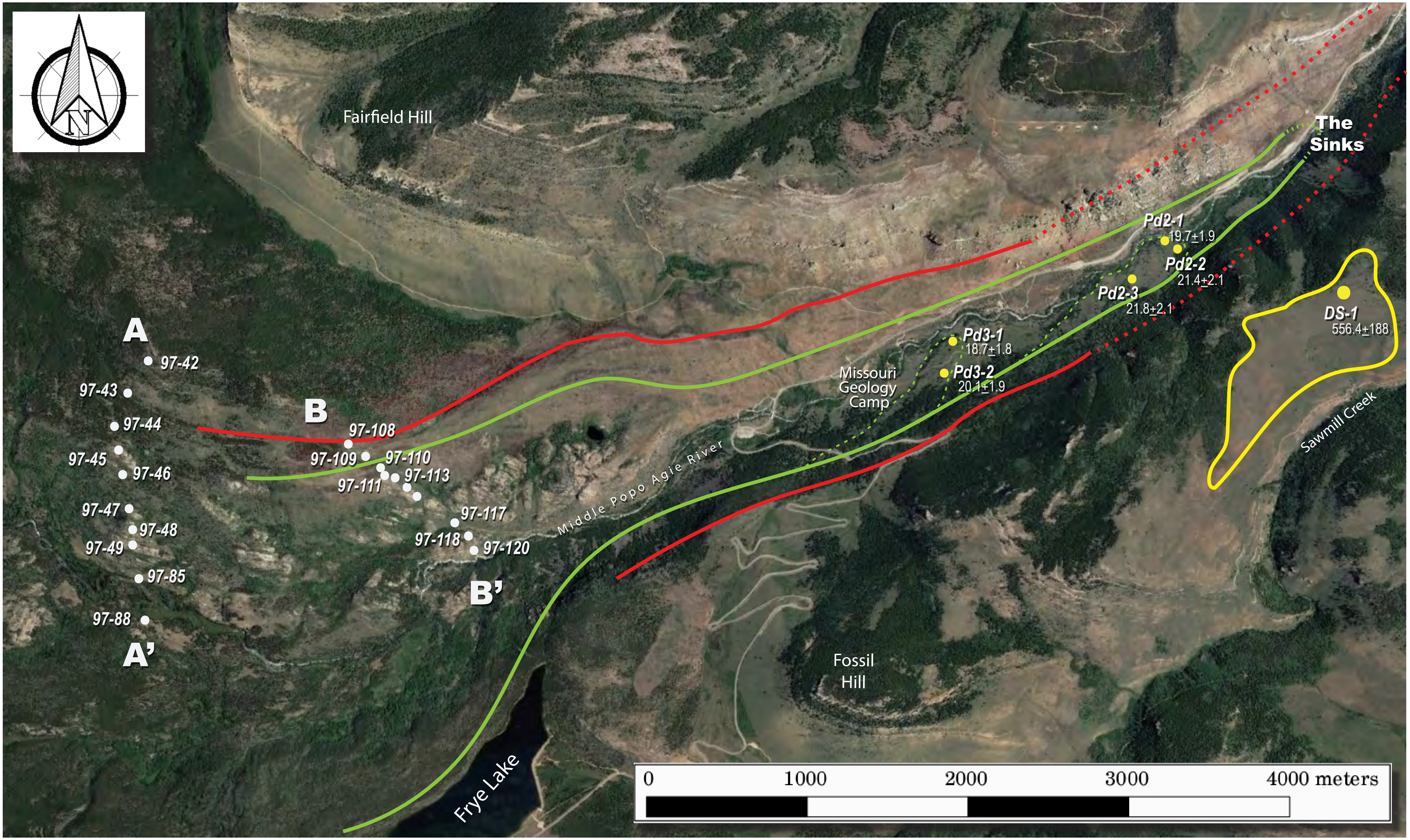
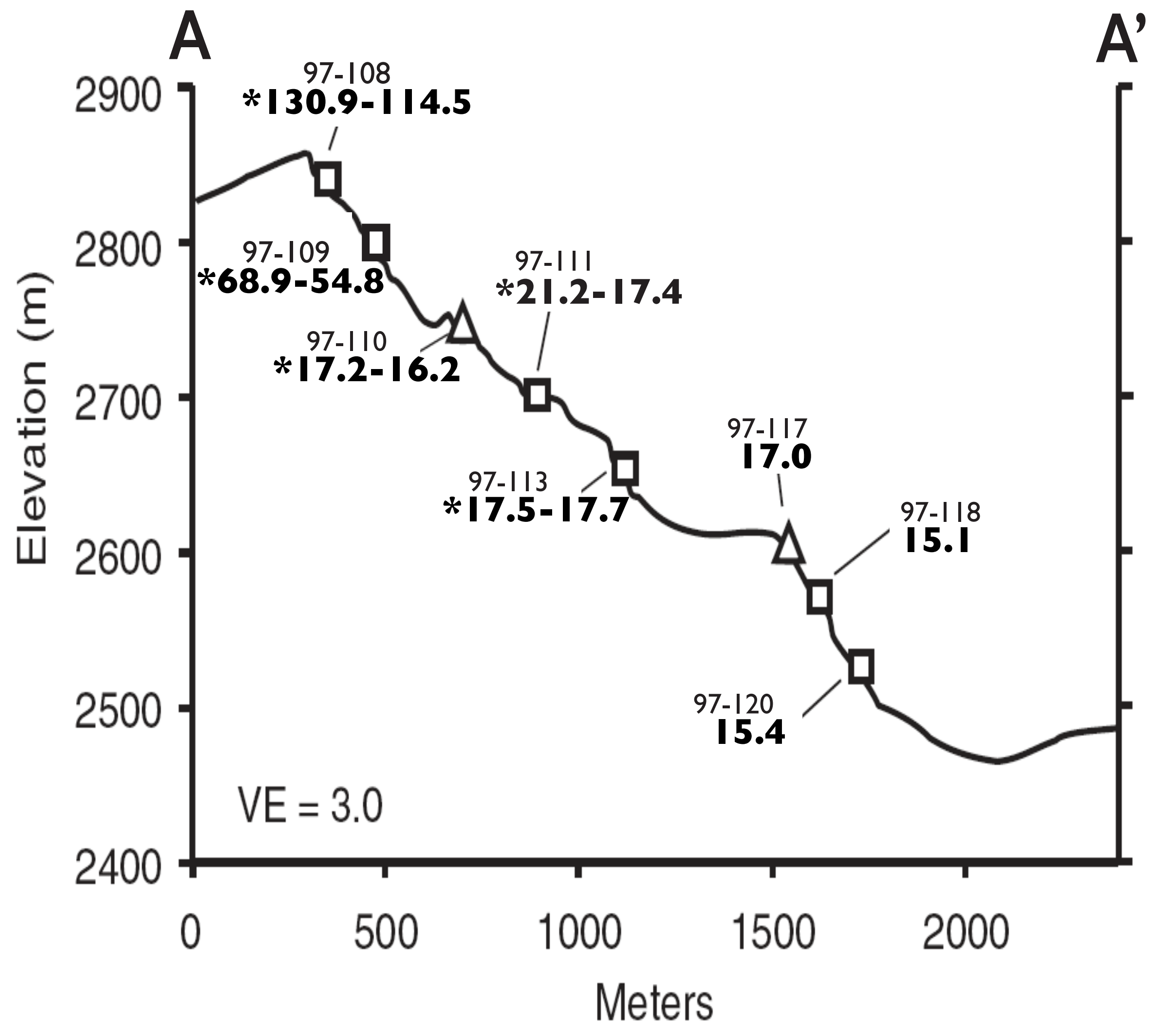
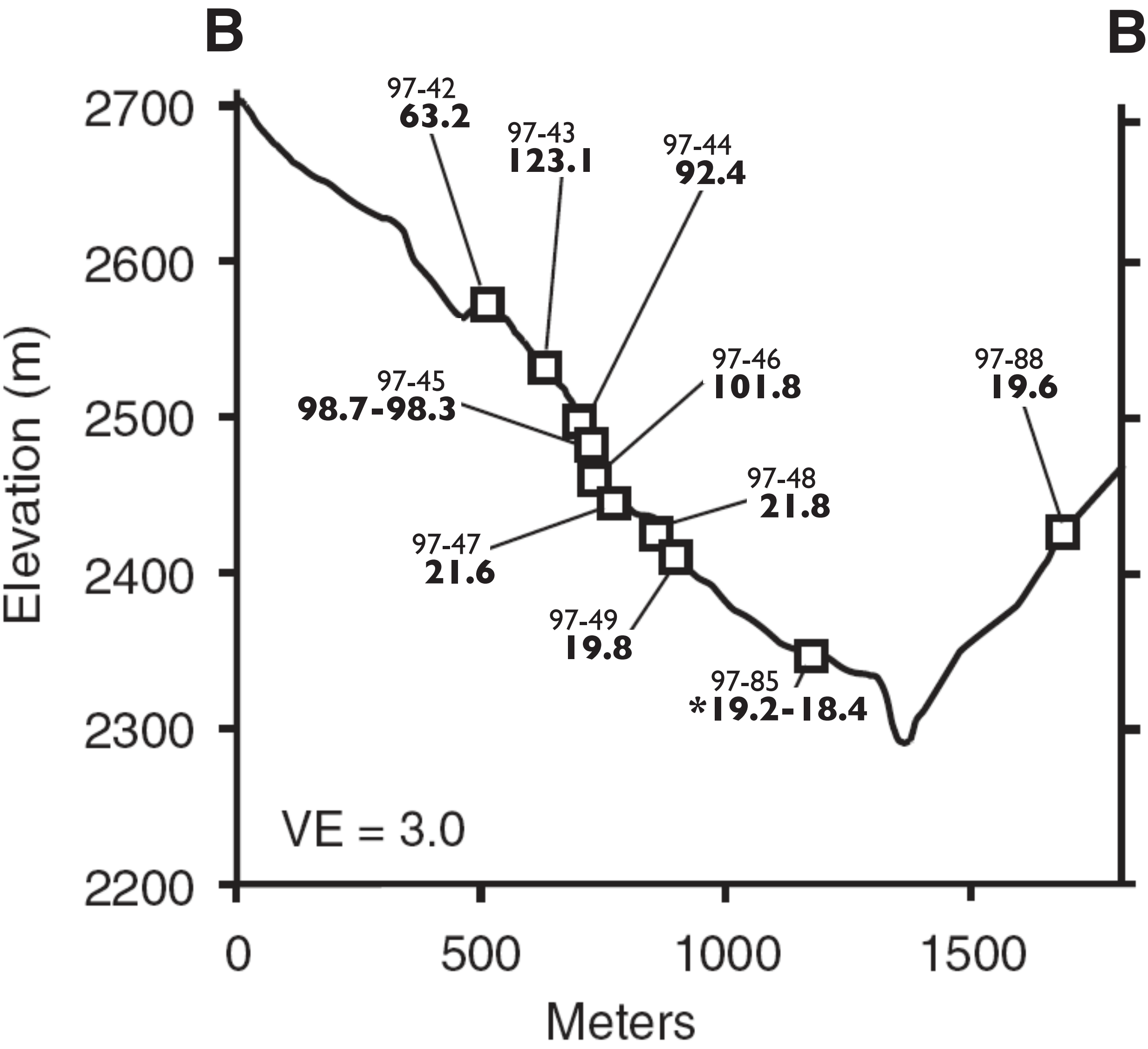


DS-1
556.4±188 ka
Sinks Canyon
The Sinks
Sawmill Creek
TM-6
>2Ma
TM-5
373.6±95 ka
TM-3
404.2±97 ka
TM-4
199.1±36 ka
TM-1
>2 Ma
TM-2
149.1±23
Table Mountain
Reentrant
ET-1
429.6±106 ka
ET-2
235.3±38 ka
HR2
HR1
N-1
126.6±16 ka
N-3
92.6±12 ka
N-2
163.3±22 ka
0 1000 2000 3000 4000 meters



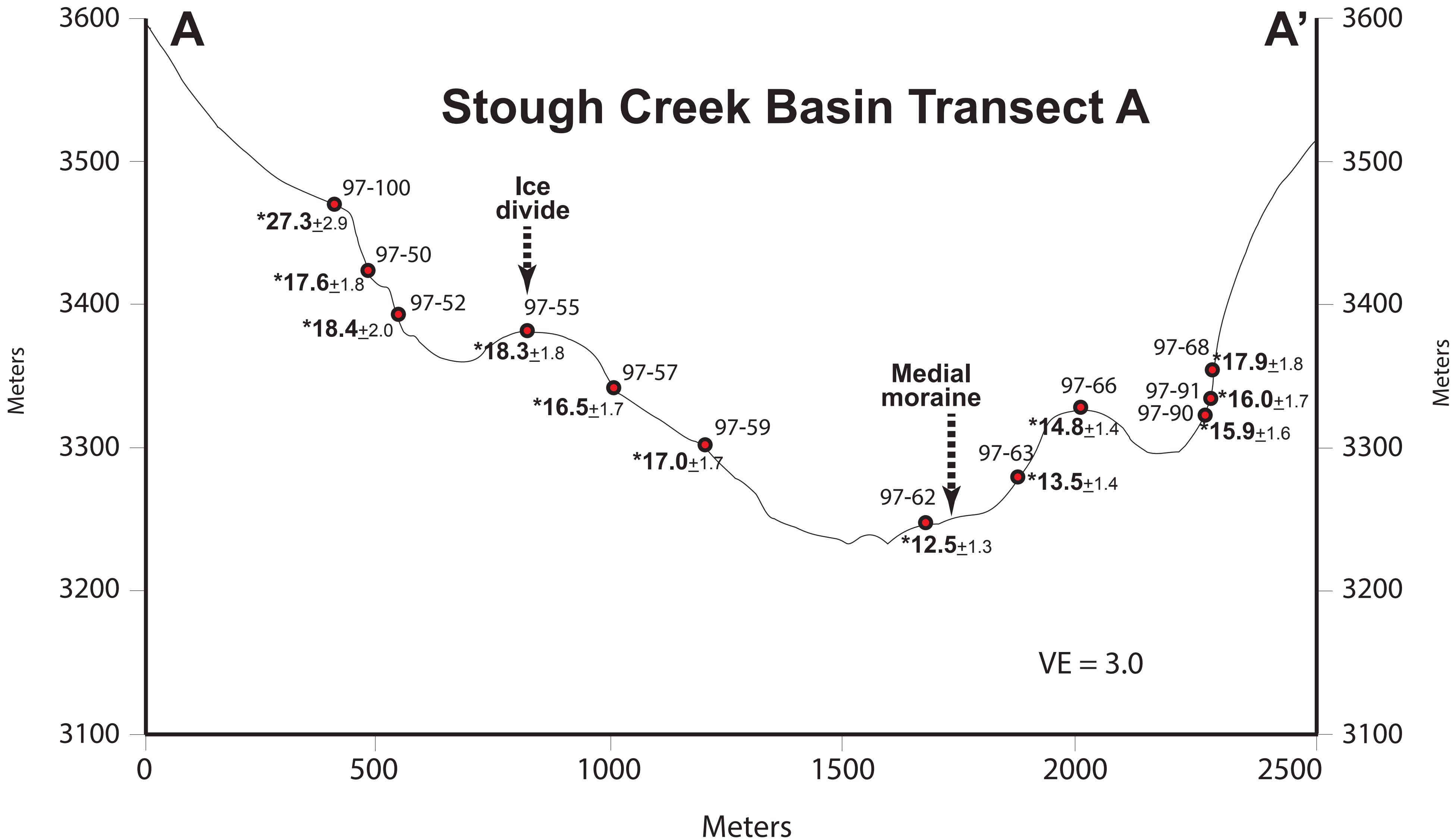
N
Fairfield Hill
The Sinks
Pd2-1
19.7±1.9
Pd2-2
21.4±2.1
Pd2-3
21.8±2.1
DS-1
556.4±188
Pd3-1
18.7±1.8
Pd3-2
20.1±1.9
Missouri Geology Camp
Sawmill Creek
A
97-42
97-43
97-44
97-45
97-46
97-47
97-48
97-49
97-85
97-88
A'
B
97-108
97-109
97-110
97-111
97-113
97-117
97-118
97-120
B'
Middle Popo Agie River
Fossil Hill
Frye Lake
0
1000
2000
3000
4000 meters


A
A'
97-108
*130.9-114.5
97-109
*68.9-54.8
97-110
*17.2-16.2
97-111
*21.2-17.4
97-113
*17.5-17.7
97-117
17.0
97-118
15.1
97-120
15.4
VE = 3.0
Elevation (m)
2900
2800
2700
2600
2500
2400
0
500
1000
1500
2000
Meters

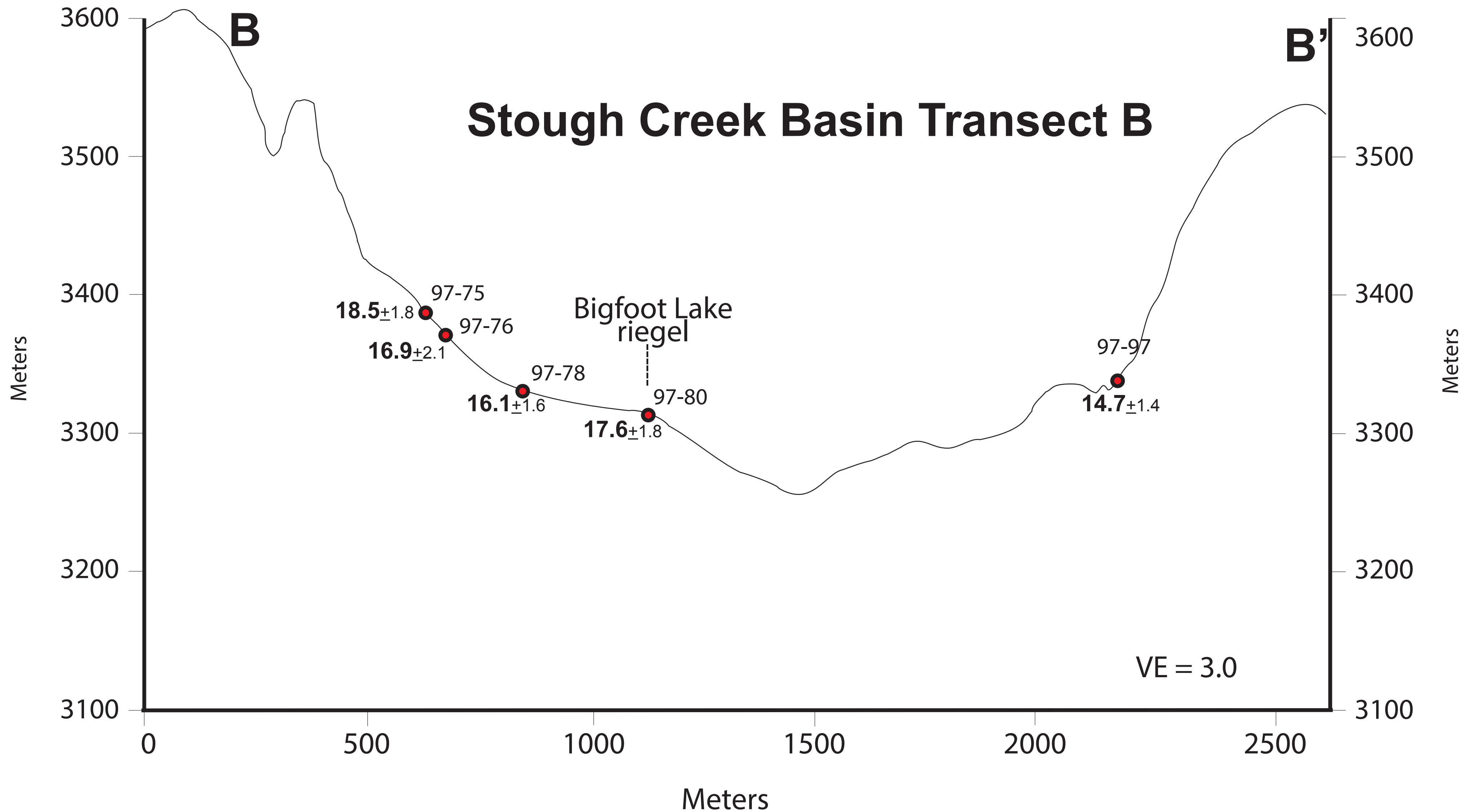


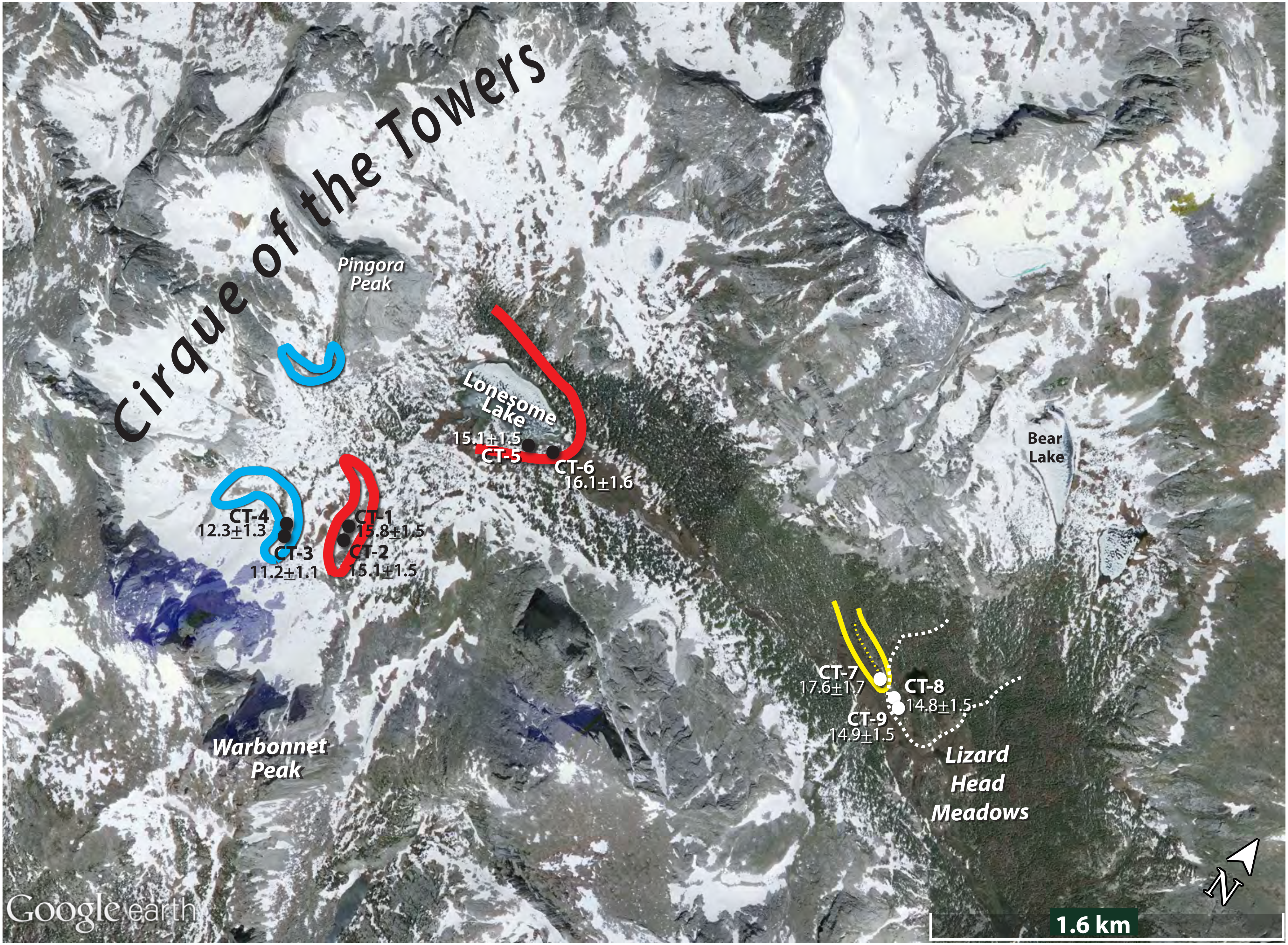
B
B'
97-42
63.2
97-43
123.1
97-44
92.4
97-45
98.7-98.3
97-46
101.8
97-47
21.6
97-48
21.8
97-49
19.8
97-85
*19.2-18.4
97-88
19.6
2700
2600
2500
2400
2300
2200
Elevation (m)
VE = 3.0
0
500
1000
1500
Meters

Roaring Fork Mountain
12.2±1.2
10.7±1.0
9.6±0.9
12.1±1.2
12.2±1.2
9.9±1.0
10.2±1.0
SCM-001-007*
SCO-001-007
12.8±1.3
14.9±1.4
15.6±1.5
14.5±1.6
15.3±1.6
13.9±1.4
13.1±1.3
13.6±1.5
12.7±1.2
AL-1
AL-3
TL-2
17.2±1.7
TL-1
13.9±1.5
Helen Lake
Bigfoot Lake
Ice Lake
Wilhelm Lake
A
B
A'
B'
97-100
97-50
97-52
97-55
97-57
97-59
97-62
97-63
97-66
97-68
97-75
97-76
97-78
97-80
97-97
97-90
15.9±1.6
97-91
16.0±1.7
1 km
1 km
Roaring Fork Mountain

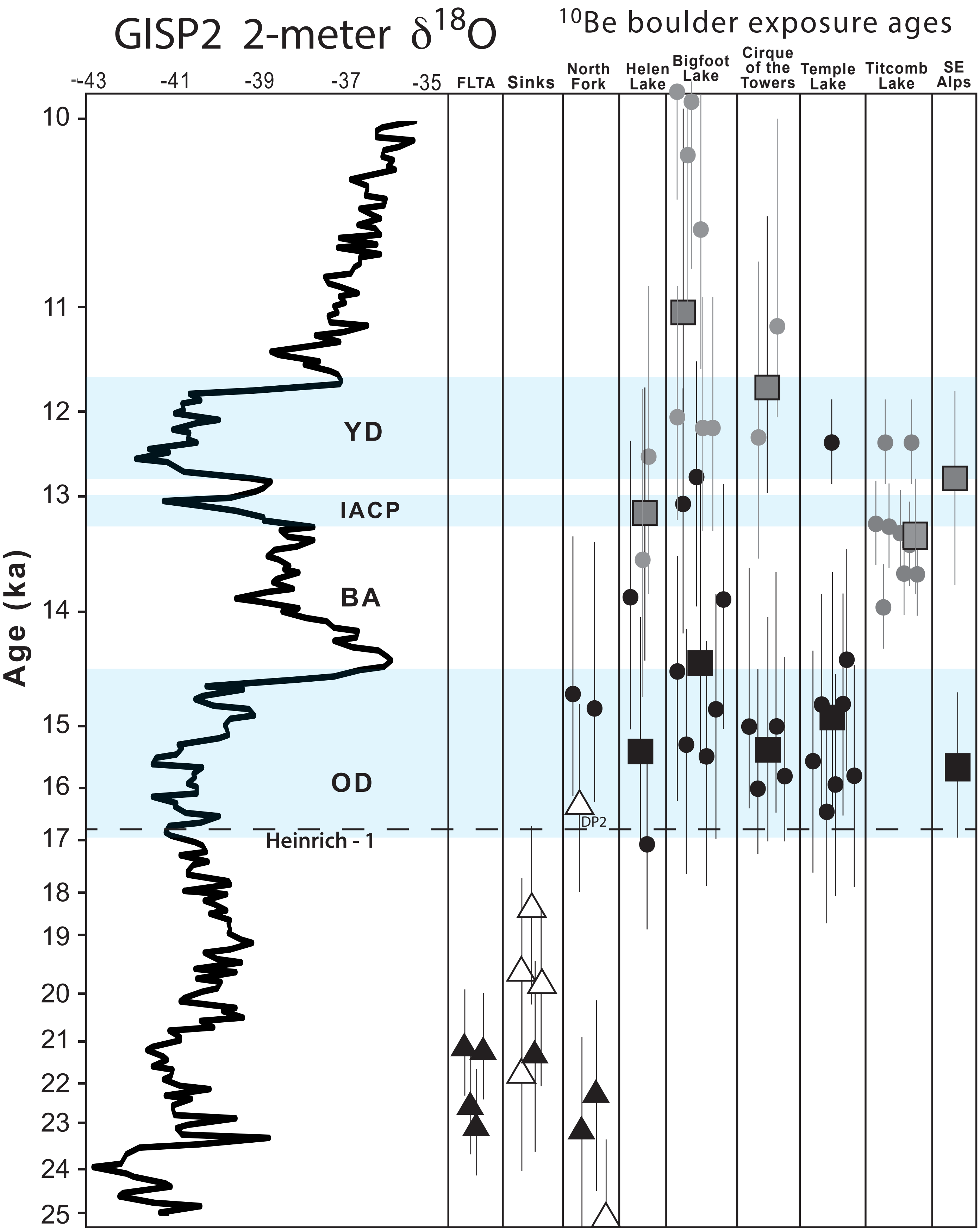


Stough Creek Basin Transect A
A
A'
3600
3500
3400
3300
3200
3100
Meters
0
500
1000
1500
2000
2500
Meters
Ice divide
Medial moraine
97-100
*27.3±2.9
97-50
*17.6±1.8
97-52
*18.4±2.0
97-55
*18.3±1.8
97-57
*16.5±1.7
97-59
*17.0±1.7
97-62
*12.5±1.3
97-63
*13.5±1.4
97-66
*14.8±1.4
97-68
*17.9±1.8
97-91
*16.0±1.7
97-90
*15.9±1.6
VE = 3.0


B
B'
Stough Creek Basin Transect B
3600
3500
3400
3300
3200
3100
Meters
18.5±1.8
97-75
97-76
16.9±2.1
Bigfoot Lake
riegel
97-78
16.1±1.6
97-80
17.6±1.8
97-97
14.7±1.4
VE = 3.0
0
500
1000
1500
2000
2500
Meters



Cirque of the Towers
Pingora Peak
Lonesome Lake
15.1±1.5
CT-5
CT-6
16.1±1.6
CT-4
12.3±1.3
CT-3
11.2±1.1
CT-1
15.8±1.5
CT-2
15.1±1.5
Bear Lake
CT-7
17.6±1.7
CT-8
14.8±1.5
CT-9
14.9±1.5
Lizard Head Meadows
Warbonnet Peak
Google earth
N
1.6 km


GISP2 2-meter δ18O
10Be boulder exposure ages
-43
-41
-39
-37
-35
FLTA
Sinks
North Fork
Helen Lake
Bigfoot Lake
Cirque of the Towers
Temple Lake
Titcomb Lake
SE Alps
Age (ka)
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
YD
IACP
BA
OD
Heinrich - 1
DP2