1 Million-year melt-presence in monotonous intermediate magma for a volcanic-plutonic
2 assemblage in the Central Andes: contrasting histories of crystal-rich and crystal-poor super-
3 sized silicic magmas

4
5 **Jason F. Kaiser¹, Shanaka de Silva², Axel K. Schmitt³, Rita Economos⁴, Mayel Sunagua⁵**

6
7 *¹Department of Physical Science, Southern Utah University, 351 West University Blvd, Cedar
8 City, Utah 84720 – jasonkaiser@suu.edu*

9 *²College of Earth Ocean and Atmospheric Sciences, Oregon State University, 104 CEOAS Admin
10 Building, Corvallis OR 97331*

11 *³Institut für Geowissenschaften Universität Heidelberg Im Neuenheimer Feld 234-236 69120
12 Heidelberg*

13 *⁴Roy M. Huffington Department of Earth Sciences, Southern Methodist University, PO Box
14 750395 Dallas, Texas 75275*

15 *⁵Calle Heroínas N° 904, Zona Miraflores La Paz-Bolivia*

16
17 **ABSTRACT**

18 The melt-present lifetime of super-sized monotonous intermediate magmas that feed
19 supereruptions and end life as granodioritic plutons is investigated using zircon
20 chronochemistry. These data add to the ongoing discussion on magma assembly rates and have
21 implications for how continental batholiths are built. Herein, we estimate ~1.1 Ma of
22 continuous melt presence before and after the climactic caldera-forming 2.89 ± 0.01 Ma (2σ)

error) Pastos Grandes Ignimbrite (PGI) supereruption ($\sim 1500 \text{ km}^3$ of magma) in the Andes of southwest Bolivia. Zircon crystallization in PGI pumice and lava from the faulted Southern Postcaldera Dome span $\sim 0.7 \text{ Ma}$ prior to the climactic eruption and formation of the eponymous caldera, whereas younger, unfaulted Postcaldera Dome lavas (termed Northern and Middle) and a granodioritic plutonic clast within the products of a Pleistocene eruption indicate a further $\sim 0.4 \text{ Ma}$ of post-climactic zircon crystallization. Bulk-rock compositions as well as zircon thermometry and geochemistry indicate the presence of homogeneous dacitic magma before and after the climactic eruption, but a trend to zircon crystallization at higher temperatures and from less evolved melts is seen for post-climactic zircon. We propose a model in which a large volume of crystal-rich dacite magma was maintained above solidus temperatures by periodic andesitic recharge that is chemically invisible in the erupted components. The climactic caldera-forming eruption vented the upper portions of the magma system where zircon was saturated. Zircon in postcaldera lavas indicate that at residual magma from this system remained locally viable for eruption at least for some time after the caldera-forming event. Subsequently, deeper “remnant” dacite magma previously outside the zone of zircon saturation rose to shallower levels to re-establish magmatic and isostatic equilibrium where zircon crystallization commenced anew, and drove more resurgent volcanism and uplift. The same magma crystallized as a granodiorite pluton which was sampled as xenoliths in much later volcanic events. Over the $\sim 1.1 \text{ Ma}$ zircon crystallization history for the PGI, postcaldera lavas and xenoliths, the melt remained in an $\sim 100\text{-}150^\circ\text{C}$ temperature window as indicated by Ti-in-zircon thermometry. Although chemical trends are consistent with zircon crystallization at variable temperatures, there is no secular cooling, but rather a thermal rejuvenation following

the 2.89 Ma PGI eruption. As such these data provide a “long and slow” temporal constraint for models for the pre-eruptive lifetimes of mushy magma in contrast to the “rapid” mobilization of crystal-poor silicic magmas, consistent with a model where that latter are incubated within the former and extracted rapidly prior to eruption. The thermal and chemical monotony of crystal-rich dacites throughout a caldera eruption cycle connotes conditions where near-eutectic melt can be maintained in near-surface magma reservoirs for an extended period of time if the subvolcanic magma reservoir is sufficiently large so that hotter and initially zircon-undersaturated magma can replenish shallow magma vented in a supereruption.

Keywords: Magmatic longevity, zircon chronochemistry, monotonous intermediates, caldera-forming eruption, plutonic clasts, Central Andes

INTRODUCTION

Large volume crystal-rich, monotonous intermediate ignimbrites, the result of climactic caldera-forming eruptions (Hildreth, 1981; Lipman, 2007), are seen as the eruptive equivalents of granodioritic plutons with the former recording the melt-present history of the latter (e.g. Bachmann et al., 2007). Whether punctuated super-sized caldera-forming eruptions with limited compositional variability are the surface expression of cordilleran batholith formation over several millions of years of incremental assembly and piecemeal construction, however, remains contested (e.g., Glazner et al., 2015). Constraints of how long a melt-present state persists in the magmas that feed super-volcanic eruptions underpin our understanding of the

“active” lifetimes of plutons, the geothermal and mineral resource potential of these systems, and any associated hazard. A handful of monotonous crystal-rich ignimbrites have been the subject of investigations that reveal, based largely on U-Pb zircon geochronology, >>100 ka pre-eruptive melt-present lifetimes (Brown and Fletcher, 1999; Costa et al., 2008, Wotzlaw et al., 2013; Kern et al., 2016). However, zircon ages for the ignimbrites only pertain to the duration of pre-climactic magma presence, whereas the significant post-climactic history of resurgent volcanism and magmatically-motivated structural uplift (Smith and Bailey, 1968; Marsh, 1984; Kennedy et al., 2012) is rarely accounted for. A more accurate quantification of the melt-present lifetime of large silicic magmatic systems thus requires that the full magmatic cycle, including the pre-climactic and post-climactic magma history culminating in solidification, be investigated. This can be achieved in systems where co-located volcanic and plutonic rocks are found, because where such a record is available, they show striking similarities in geochemistry and petrology that suggest a cogenetic relation (e.g. Lipman, 2007). With the development of techniques that allow coupled chronological and chemical measurements of zircon, we are now in a position to also estimate melt-present lifetimes for volcanic systems where plutons are not exposed, albeit in favorable cases accessible as xenoliths.

Herein, we apply zircon chronochemistry, the combination of U-Pb geochronology with indicators for geochemical evolution (e.g., Zr/Hf, Yb/Gd, Eu/Eu*, U/Th) and model magmatic temperatures for zircon (e.g., through zircon saturation and Ti-in-zircon thermometry), to volcanic rocks and plutonic clasts from the Pastos Grandes caldera in southwest Bolivia. The erupted rocks define a “monotonous intermediate” suite that records the pre- and post-climactic history of a super-sized caldera-related magma system. Here zircon records a quasi-

continuous >1 Ma history of thermal, compositional, and textural evolution in a shallow crystal-rich dacitic crustal magma reservoir that is interpreted to have evolved into a regional composite batholith.

GEOLOGIC BACKGROUND

The Pastos Grandes Caldera Complex (PGCC) in southwestern Bolivia (Fig. 1) is part of the Altiplano-Puna Volcanic Complex (APVC), an ignimbrite plateau active from ~10 Ma to Recent (<100 ka) (de Silva, 1989; de Silva et al., 2006, Salisbury, et al., 2011). The calderas of the APVC overlie a ~500,000 km³ low-velocity zone, the Altiplano-Puna Magma Body (APMB) that is interpreted as the magmatic feeder zone of a cogenetic batholith (Ward et al., 2014; Burns et al., 2015). The earliest PGCC caldera cycle resulted from the 5.45 ± 0.02 Ma eruption of the 1200 km³ dense rock equivalent (DRE) Chuhuilla Ignimbrite (Salisbury et al., 2011). A second climactic eruption at 2.89 ± 0.01 Ma resulted in the 1500 km³ (DRE) Pastos Grandes Ignimbrite (PGI) and produced the eponymous ~40 × 25 km caldera nested within the larger Chuhuilla caldera (Fig. 1). Following the eruption of the PGI and the formation of the caldera was a series of effusive eruptions resulting in lava domes on the caldera floor. Two types of undated, but structurally distinct post-climactic domes (PCD) are present. The older, Southern PCD is faulted by resurgent uplift of the caldera floor. By contrast, younger post-climactic volcanic activity resulting in the Mid and North PCD are unaffected by resurgence, indicating they erupted after resurgent uplift. The distinct ~20 km diameter, 1 km-high resurgent dome and its surrounding post-climactic lava flows and domes define the footprint of magmatic and structural resurgence for the Pastos Grandes supereruption. The plutonic part of the Pastos Grandes system is

represented by plutonic clasts erupted as ejecta during the much younger (~85-94 ka) Chascon Runtu-Jarita lava-complex episode (Watts et al., 1999). The few recovered clasts were between 5-10 cm in diameter, and only one of the clasts was used to sample zircons for this study.

PGI pumice clasts are crystal-rich, calc-alkaline dacites between 67 wt% and 70 wt% SiO₂ which contain ~40% phenocrysts (vesicle free) of plagioclase (maximum diameter 3 mm), quartz, biotite, amphibole, and rare sanidine, set in a vesicular, high-Si rhyolite (76 – 77% SiO₂) matrix glass. Accessory phases include apatite, titanite, Fe-Ti oxides, and zircon. The post-climactic domes and plutonic clasts have bulk rock compositions closely overlapping those of the PGI (Fig. 2). The post-climactic lavas are similar in mineralogy to the PGI pumice but are denser and coarser (maximum diameter 5 mm for plagioclase and quartz phenocrysts) and with higher phenocryst content (up to 50% based on approximations from petrography). The groundmass of the post-climactic lavas is rich in microlites. The plutonic clasts are granodiorite and have a phaneritic texture with dominant plagioclase, potassium feldspar, biotite, amphibole, and quartz (Fig. 2). Thus, the cogenetic suite of volcanic and plutonic rocks shares a general family resemblance in mineralogy and whole rock chemistry with a trend of increasing textural coarseness (maturity) from pumice to lava to granodiorite (Fig 2). The chemical similarities extend to the zircon populations as we elaborate below.

METHODS

Sample Preparation

Zircon crystals were recovered by heavy liquid separation from the 100-250 µm sieved fraction of crushed rock samples. Grains were mounted in epoxy and polished to intersect

crystal interiors. Prior to ion probe analyses, each zircon crystal was examined for internal zoning by cathodoluminescence (CL) imaging using a Leo 1430VP scanning electron microscope equipped with an Oxford Mini-CL detector at University of California Los Angeles (UCLA) (Fig. 3). A small subset of representative zircon crystals is shown in the CL images, which were used to target core and rim locations for ion probe analysis with a preference for CL homogenous domains to minimize averaging of multiple growth zones in the zircon crystals.

Analytical Procedures

Ion microprobe analyses were performed using the CAMECA ims 1270 ion microprobe at UCLA. A mass-filtered $^{16}\text{O}^-$ primary beam with an intensity of 15-20 nA was focused to a size of 25-30 μm . For U-Pb analyses, secondary ions were extracted at 10 kV, a mass resolution power $m/\Delta m = 4500$ at 10% of the peak height, and an energy bandpass of 50 eV. Replicate analysis of reference zircon AS3 (Paces and Miller, 1993) yielded a $^{206}\text{Pb}/^{238}\text{U}$ age reproducibility of <3%. U-Pb geochronology ion probe craters were subsequently used for trace element analyses after re-imaging the crystals by CL, and repolishing them to remove all but a small vestige of the U-Pb crater. The $^{206}\text{Pb}/^{238}\text{U}$ ages were determined after a ^{207}Pb -based common Pb correction (common $^{207}\text{Pb}/^{206}\text{Pb} = 0.8283$) and an age correction for ^{230}Th deficit based on the calculated partitioning from measured Th/U in zircon and Th/U in the melt (2.16). The trace element spot was then placed directly onto the U-Pb spot using a $\sim 15\text{-}20$ nA $^{16}\text{O}^-$ primary ion beam. High energy (+100 eV) secondary ions were analyzed at intermediate mass resolution ($m/\Delta m = 2000$) using energy-filtering to suppress molecular interferences. Remaining oxide interferences (light rare earth element REE oxides overlapping onto heavy REE) were subtracted

using atomic to oxide ratios determined from analysis of synthetic glasses selectively doped with REE to generate interference-free patterns. Trace element relative sensitivity factors were calibrated using NIST610 glass (Pearce et al., 1997). Replicate analyses of zircon standard 91500 agree within <20% with values reported in Liu et al. (2010) for Ti, Zr/Hf and REE with the exception of La (0.03 ppm vs. 0.005 ppm).

Treatment of U-Pb Data

A total of 186 U-Pb zircon analyses for seven samples representing five units are included here; for individual units the number of grains analyzed are as follows: PGI n = 58, South PCD n = 31, Mid PDC n = 22, North PCD n = 22, and Gran n = 34. Mean square of weighted deviate (MSWD) values for common-Pb and ^{230}Th disequilibrium-corrected $^{206}\text{Pb}/^{238}\text{U}$ ages exceed the high-95% critical limit for a single population (Mahon, 1996) for all samples (PGI MSWD = 5.71; Gran = 4.54; North PCD = 2.63) except for South PCD (MSWD = 1.43) and Middle PCD (MSWD = 1.55). Probability density functions for individual populations with high MSWD values are skewed towards older ages. We interpret these observations as the result of zircon crystallization over durations which exceed analytical resolution. Because of this, averages do not adequately represent the complexities of the age populations. To refine our interpretation of magma presence through zircon chronology, we applied the age unmixing algorithm of Sambridge and Compston (1994) as implemented in Isoplot Version 3.71 (Ludwig, 2008). PGI, NPCD, SPCD, and the plutonic clast sample (Gran) populations were deconvolved assuming two normally-distributed sub-populations in the unmixing algorithm; models with more than two populations failed to decrease relative misfit values. In all cases, the dominant (at 51-87%) sub-

populations yielded younger ages and the older sub-population was generally less abundant. These sub-populations may or may not be significant, especially when individual ages represent a continuum as shown here (Fig. 4), but the corresponding age modes provide a minimum estimate for zircon crystallization when accounting the dispersion of the data within each population solely to analytical uncertainties. This data treatment also shows that a large portion of zircon crystals formed close to the eruption age.

Strong similarities in the zircon age distributions for certain samples lead us to categorize the data into two major groups: the PGI pumice and South PCD zircon crystals make up the older, pre-climactic stage, while zircon crystals from the Mid PCD, North PCD, and Gran make up the younger, post-climactic stage. The case of the South PCD is peculiar: while the dome clearly represents post-caldera volcanic activity, zircons from the South PCD samples exclusively yield pre-climactic stage ages. The individual and categorized age distributions are described together with their trace element characteristics below.

ZIRCON AGES AND TRACE ELEMENT COMPOSITIONS

A first order observation for the zircon data is that in all units clearly xenocrystic ages (i.e. 10.0, 17.2, and 147 Ma) represent only ~2% of the population, and are exclusively found in crystal interiors (Fig. 3). The remaining 98% of analyses for the two main age groups are distributed around the PGI eruption age (Fig. 4). Zircons from the PGI and South PCD yield ages with an older mode at 3.19 ± 0.05 Ma and 3.11 ± 0.03 Ma respectively (age population errors also 1σ). Figure 4 shows that these two populations overlap defining the peak of zircon crystallization for the pre-climactic stage at 3.14 ± 0.04 Ma for the younger sub-population. This

age represents the highest probability for zircon crystallization, or magmatic age, for the pre-climactic stage, and significantly predates the sanidine $^{40}\text{Ar}/^{39}\text{Ar}$ climactic eruption age (2.89 ± 0.01 Ma; Salisbury et al., 2011). The older zircon sub-population for the pre-climactic stage averages 3.63 ± 0.06 Ma (shown as small shoulder on curve in Figure 4). The difference between the PGI eruption age and the older subpopulation average age computes to a model pre-eruptive zircon crystallization duration of 0.74 ± 0.06 Ma. The post-climactic stage zircons (samples Mid PCD, North PCD and Gran) define two subequal age peaks at 2.52 ± 0.03 Ma and 2.75 ± 0.03 Ma with hardly any overlap with the zircon ages from the pre-climactic stage. Taking the difference between the oldest pre-climactic model population at 3.61 ± 0.02 Ma to the post-climactic population at 2.52 ± 0.03 Ma yields a nominal duration of zircon crystallization of 1.10 ± 0.07 Ma for the PGCC which may extend from the earliest zircon crystallization during the initial build-up stage to the terminal solidification represented by the granodiorite. Alternatively, if we only consider the difference between the pre-climactic and post-climactic peak autocryst ages to define the time period from magma accumulation to final pluton solidification (dismissing the antecrystic zircon population of the pre-climactic stage as unrelated to the cataclysmic PGI magma), the time interval between the major zircon crystallization peaks is 0.62 ± 0.05 Ma. Because the analytical uncertainties of the individual zircon ages permit longer crystallization durations, both durations should be combined to reach a minimum for the longevity of zircon crystallization in the PGCC magma system.

Despite these clear age differences, zircons in pre- and post-climactic magma populations independently show nearly identical trace element signatures and trends. Selected zircon trace element ratios (Fig. 5) reveal that zircon crystallized under very similar temperature

221 and compositional conditions during pre- and post-climactic stages in the shallow crust.

222 Overlapping pre- and post-climactic Ti vs Zr/Hf trends indicate similar thermochemical

223 conditions for zircon crystallization (Fig. 5a). Zr/Hf is used to effectively isolate the fractionating

224 characteristics of zircon as no other phase significantly fractionates Zr from Hf (Claiborne et al.,

225 2006). Trace element ratios vs Zr/Hf (Fig.5b-d) in the context of a semi-quantitative chemical

226 model (supplemental material – zircsupp1) corroborates expected behavior based on published

227 partitioning values (e.g. Wotzlav et al., 2013; Blundy and Wood, 2003, Linnen and Keppler,

228 2002; Sano et al., 2002; Mahood and Hildreth, 1983; Luhr and Carmichael, 1980) and supports

229 co-crystallization of hornblende, titanite, and allanite with zircon (e.g., Claiborne et al., 2006;

230 Reid et al., 2011). These trends suggest that crystallization of zircon and titanite is to a first

231 order temperature-dependent (Green and Pearson, 1986; Boehnke et al., 2013). Whereas the

232 ranges in temperature and chemistry overlap between pre- and post-climactic zircon, post-

233 climactic zircons spread out to slightly higher temperatures and less evolved compositions (Figs.

234 6 and 7). The changes in trace element concentrations reflect melt evolution, following a

235 decreasing temperature trend as crystals grow. By using the calibration of Ferry and Watson

236 (2007), and a constant Ti activity of 0.7, a model Ti-in-zircon temperature range was calculated

237 at ~200 °C (Table 1 – full chemical data are listed in the online supplemental material

238 zircsupp2). This assumes that activity of TiO_2 was constant or increased with decreasing

239 temperature when buffered by an equilibrium Fe-Ti oxide assemblage (Ghiorso and Gualda,

240 2013). It should be noted that a larger temperature range could have been calculated by

241 changing activities, and that 0.7 was chosen as a typical value for silicic volcanic rocks lacking

242 rutile (Ghiorso and Gualda, 2013). The temperature range is taken as another line of evidence

for similar magmatic conditions in the pre- and post-climactic stages. Given this temperature range, and the ca. 1 Ma longevity of zircon crystallization, we compute a very slow cooling rate of $2 \times 10^{-4} \text{ }^{\circ}\text{C/a}$, although this is a maximum estimate for the entire system where the zircon data do not reveal any secular cooling trends. The trends of Figures 5 and 6 suggest largely invariant chemical and thermal conditions in the pre- and post-climactic magma stages over the ca. 1 Ma time span of zircon crystallization.

DISCUSSION

We interpret the zircon ages in the PGCC to indicate the presence of Zr-saturated melt where the age range defined by zircon is a measure of the minimum “magmatic” lifetime of the system. This magmatic lifetime is a minimum for the time period ranging from pre-climactic magma accumulation to post-climactic resurgence and pluton solidification.

Magmatic lifetime of the PGCC magma

We find statistically resolvable zircon age differences in individual samples that exceed purely random analytical noise (indicated by the high MSWD values for the population). Based on the age components from the unmixing model, four major episodes of zircon crystallization can be discerned at ca. 3.6, 3.1, 2.8, and 2.5 Ma, each within an uncertainty of $\pm 0.03\text{-}0.06$ Ma (the eruption age being 2.89 Ma). Alternatively, if the geologic age variation was equivalent to the analytical uncertainty, the data could represent continuous crystallization over timescales of >1 Ma, with the second, post-climactic, stage of zircon crystallization commencing virtually immediately after the climactic eruption. Therefore, regardless if we use the age continuum of

individual analyses or the model peaks, we conclude that the zircon crystallization duration (melt presence) between pre-eruptive accumulation in the PGI and terminal solidification seen in the post-climactic stage samples and plutonic clasts extended over 1.1 Ma at least. Protracted and continuous zircon crystallization, rather than individual short-lived episodes of crystallization, are consistent with the progression of textural maturation we see in the samples and the evidence of a minor antecryst population in the post-climactic population. Piecemeal assembly over much larger timescales than 1.1 Ma is unsupported by the data, despite the regional continuity of magmatism over nearly 10 Ma. In this context, the absence of zircon in the Pastos Grandes cycle from the precursor Chuhuilla Ignimbrite cycle at 5.45 ± 0.02 Ma is intriguing. Not only is the Chuhuilla system too old to be reasonably considered part of a magmatic continuum leading to the PGI eruption, but the lack of Chuhuilla zircon ages suggests that resorption of zircon from pre-existing intrusions in the build-up phase of a large magma reservoir was very efficient (cf. Glazner et al., 2015).

It is intriguing that identical zircon populations between PGI and post-climactic eruptions only exist for the Southern PCD, whereas most zircons in the post-climactic Northern and Middle PCD and granodiorites are autocrystic, despite the significant volume contrasts between the preceding cataclysmic eruption and the subsequent post-climactic dome eruptions. This suggests either that eruptible volumes of the pre-climactic magma may have been virtually exhausted or that any remnant magma was heated to above zircon saturation temperatures resulting in resorption of pre-existing zircon. We eschew the former because post-caldera eruptions and the plutonic clasts are virtually identical to the PGI and the textural progression would be consistent with continued “ripening” of the pre-climactic PGI-forming

287 magma. While this could have been a small volume, we interpret the extremely slow secular
288 cooling rate revealed by the post-climactic Ti-in-zircon thermometer to require a substantial
289 body of several 100 km³ at least. The alternative, remnant magma becoming reheated, is a
290 possibility, but the lack of resorption of other mineral phases and again, the textural
291 progression, is inconsistent with this transient up-temperature evolution. Moreover,
292 cathodoluminescence images of zircons (Fig. 3) show an undisturbed oscillatory zoning pattern
293 with obvious truncations being scarce, indicating continuous, or nearly continuous
294 crystallization. Thus, we propose that the magmatic history sampled by the PGCC eruptions
295 may be best understood in terms of a single magma that was divided into two zones
296 distinguished solely by zircon stability but simultaneously crystallizing major phases. Initially the
297 upper part was saturated in zircon, whereas temperatures in the lower parts were above zircon
298 saturation. After the caldera-forming eruption and extrusion of the Southern PCD, the remnant
299 magma and its crystal cargo rose as a result of the loss of the confining effect of a coherent roof
300 to re-establish magmastatic and isostatic equilibrium (e.g. de Silva et al., 2015). This rise drove
301 resurgent uplift and resulted in cooling below zircon saturation. As this “resurgent” intrusion
302 moved into the zircon stability field, new zircon crystallized. At least some of magma remaining
303 from the caldera forming eruption mixed in, explaining the few antecrysts from the pre-
304 climactic population (Fig. 6). The Southern post-climactic dome, with its cargo of pre-climactic
305 zircons, sampled only remnant pre-climactic magma, while the Northern and Middle PCD,
306 containing only post-climactic aged zircon, mark effusive leaks from this resurgent intrusion.
307 The distinct zircon populations in otherwise chemically homogeneous magma suggest that the
308 post-climactic zircon population followed the same crystallization sequence as during the pre-

climactic magma stage: a result of lithostatic, thermal, and bulk chemical environments of the system remaining essentially unchanged. The textural maturity of the domes implies a more advanced stage of crystallization than the climactic magma, with final solidification being recorded in the holocrystalline granodiorite clasts where zircon ages fall entirely within a ca. 0.37 Ma time window after the climactic eruption despite the much younger, late Pleistocene eruption age of the clasts.

Implications for super-sized magma systems

The Pastos Grandes ignimbrite is an example of the 'monotonous intermediate' genre, the few well characterized examples of which all demonstrate >>100 ka pre-eruptive melt-present histories (Whakamaru Ignimbrite - Brown and Fletcher, 1999; La Pacana Ignimbrite - Schmitt et al., 2002; Bachmann et al., 2007; Cerro Galan Ignimbrite - Folkes et al., 2011; Fish Canyon Tuff - Wotzlaw et al., 2013). These estimates are orders of magnitude longer than the pre-eruptive histories of super-sized, crystal-poor rhyolites of the Taupo Volcanic Zone, Long Valley caldera, and Yellowstone where assembly to eruption time scales of millennia to decades have been suggested (Allan et al., 2013; Wotzlaw et al., 2014; Pamukcu et al., 2015). This dichotomy reflects that silicic magmas may undergo different stages in their history. The first is the extended and piecemeal process of generation, accumulation, crystallization, recharge and assimilation that leads to a mushy magma with an interstitial melt of ternary eutectic composition (Hildreth, 2004). In a second stage, the interstitial melt is very rapidly expelled and segregated into a crystal-poor rhyolite body prior to eruption (e.g. Lindsay et al., 2001; Bachman and Bergantz, 2004). In this context, the PGI, and most other monotonous

intermediates, only record the long and slow first stage of development and conditions for segregation of the residual melt into a separate pre-eruptive body did not develop – a notable exception being the Atana/Toconao magma system (Lindsay et al., 2001; Schmitt et al, 2002).

The protracted longevity of >1 Ma for the PGI is consistent with the understanding that large-volume monotonous intermediate magma systems assemble and evolve over long time scales at rates that must ensure their eruptibility (Smith, 1979; Reid, 2003; Gelman and Bachmann, 2013). Such magmatic longevity is attributed to their development in a thermally mature crust, prepared by precursory magmatism and an “open” architecture where recharge provides regular thermal energy input and promotes feedbacks that enhance longevity (de Silva and Gregg, 2014). Evidence from the PGI is that recharge is “cryptic” in the sense that it does not appear to result in major chemical and thermal excursions in the upper eruptible parts of the system. At deeper, mid-crustal levels, recharge by andesitic or more mafic parental magmas is invoked as the crucial heat engine that keeps the dacitic upper crustal magmas simmering (e.g. Bachman and Bergantz, 2008; Burns et al., 2015). Albeit volumetrically very minor, the presence of andesitic and mixed pumices and andesitic enclaves is direct evidence for a prominent role of andesitic recharge which is largely cryptic because of the buffering mass of a crystal-rich mush containing evolved partial melt. Such andesitic and mixed pumice was found in almost every monotonous intermediate ignimbrite within the APVC, including PGI. The dominance of crystal-rich (40 to 50% phenocrysts) dacite of 65 to 70% SiO₂ with eruption temperatures ranging from 750 to 850 °C and minimal secular cooling indicated by Ti-in-zircon thermometry, suggests large magma reservoirs where the upper erupted parts are buffered thermally and compositionally. In combination with the high Si-rhyolite matrix glass (>76% SiO₂)

of these crystal-rich pumices, these reservoir characteristics implicate that magma evolution can stall at the eutectic for extended periods of time (e.g. Huber et al., 2010), with melt presence being maintained by recharge or simply because such crystal-rich magmas are insensitive to changes in the enthalpy content of the system (Caricchi and Blundy, 2015).

CONCLUSIONS

Chronochemical data suggest prolonged magma assembly, punctuated by large-volume eruptions, which reflects ongoing discussions of protracted continental batholith construction. Zircon crystallization ages and chemistry from pumice, lava, and plutonic clasts associated with the 2.89 Ma Pastos Grandes (PGI) caldera-forming eruption reveal protracted zircon crystallization over ca. 1 Ma durations. These data connote a single, long-lived monotonous intermediate magma that was chemically buffered but partitioned by zircon stability where two dominant populations of zircons were successively formed when initially saturated magma vented during a climactic eruption, and was replaced by chemically similar but undersaturated dacite magma that subsequently achieved zircon saturation during the post-climactic stage. This interpretation is supported by the chemical homogeneity and progressive textural maturity from the ignimbrite, through the Post-Climactic Domes, to the cogenetic granodiorite clasts. Together, these magmatic stages account not only for the pre-eruptive evolution of magma that was interrupted by climactic eruption, but also for an equally long post-climactic stage magmatism that culminates in pluton solidification within ca. 0.37 ka after the cataclysmic eruption. This time frame is longer than those reported for the pre-eruptive histories for crystal-poor rhyolites as those are proposed here to reflect only a short melt segregation period

just prior to eruption. Instead, this protracted “melt present” lifetime deduced from zircon crystallization ages provides a firm temporal constraint for the accumulation, crystallization, recharge and assimilation of mushy monotonous silicic magma and reflects the occupation of a petrological “sweet spot” where limited thermal and compositional evolution takes place.

ACKNOWLEDGMENTS

This work was enabled by support from National Science Foundation grants EAR 0838536 and 0908324. The ion microprobe facility at the University of California– Los Angeles (UCLA) is partly supported by a grant from the Instrumentation and Facilities Program, Division of Earth Sciences, National Science Foundation.

REFERENCES CITED

- Allan, A. S. R., Morgan, D. J., Wilson, C. J. N., & Millet, M.-A., 2013. From mush to eruption in centuries: assembly of the super-sized Oruanui magma body. *Cont. Min. Pet.* 166, 143–164.
- Bachmann, O., Miller, C.F., de Silva, S.L., 2007. The volcanic-plutonic connection as a stage for understanding crustal magmatism. *J. Volc. Geotherm. Res.* 167, 1-23.
- Bachmann, O., Bergantz, G. W., 2004. On the Origin of Crystal-poor Rhyolites: Extracted from Batholithic Crystal Mushes. *J. Petrol.* 45, 1565–1582.
- Bachmann, O., Bergantz, G. W. 2008. Deciphering Magma Chamber Dynamics from Styles of Compositional Zoning in Large Silicic Ash Flow Sheets. *Rev. Min. Geochem.* 69, 651–674.
- Blundy, J., and Wood, B., 2003. Mineral-melt partitioning of uranium, thorium and their daughters. *Rev. Min. Geochem.* 52, no. 1 59-123
- Boehnke, P., Watson, E. B., Trail, D., Harrison, T. M., Schmitt, A. K., 2013. Zircon saturation re-revisited. *Chem. Geol.* 351, 324-334.

- 408
409 Brown, S. J., Fletcher, I. R., 1999. SHRIMP U-Pb dating of the preeruption growth history of
410 zircon from the 340 ka Whakamaru Ignimbrite, New Zealand: Evidence for >250 k.y.
411 magma residence times. *Geology*, 27, 1035–1038.
412
- 413 Burns, D.H., de Silva, S.L., Tepley III, F., Schmitt, A.K., Loewen, M.W., 2015. Recording the
414 transition from flare-up to steady-state arc magmatism at the Purico-Chascon volcanic
415 complex, northern Chile. *EPSL*. 422, 75-86.
416
- 417 Caricchi, L., Blundy, J., 2015. Experimental petrology of monotonous intermediate magmas.
418 *Geol. Soc. Lond. Spec. Pub.* 422, 105-130
419
- 420 Claiborne, Lowery L., Miller, C.F., Walker, B.A., Wooden, J.L., Mazdab, F.K. Bea, F., 2006.
421 Tracking magmatic processes through Zr/Hf ratios in rocks and Hf and Ti zoning in
422 zircons: An example from the Spirit Mountain batholith, Nevada. *Min.*
423 *Mag.* 70, 517-543.
424
- 425 Costa, F., Dohmen, R., Chakraborty, S., 2008. Time Scales of Magmatic Processes from Modeling
426 the Zoning Patterns of Crystals. *Rev. Min. Geochem.* 69, 545–594.
427
- 428 de Silva, S., 1989. Altiplano-Puna Volcanic Complex of the Central Andes. *Geology* 17, 1102-
429 1106.
430
- 431 de Silva, S., Zandt, G., Trumbull, R., Viramonte, J. G., Salas, G., Jiménez, N., 2006. Large
432 ignimbrite eruptions and volcano-tectonic depressions in the Central Andes: a
433 thermomechanical perspective. *Geol. Soc. Lond. Spec. Pub.* 269, 47–63.
434
- 435 de Silva, S.L., Gregg, P.M., 2014. Thermomechanical feedbacks in magmatic systems:
436 Implications for growth, longevity, and evolution of large caldera-forming magma
437 reservoirs and their supereruptions. *J. Volc. Geotherm. Res.* 282, 77-91.
438
- 439 Ferry J.M., Watson E.B., 2007. New thermodynamic models and revised calibrations for the Ti-
440 in-zircon and Zr-in-rutile thermometers. *Contrib. Mineral. Petrol.* 154, 429–437.
441
- 442 Folkes, C. B., de Silva, S. L., Schmitt, A. K., Cas, R. A. F., 2011. A reconnaissance of U-Pb zircon
443 ages in the Cerro Galán system, NW Argentina: Prolonged magma residence, crystal
444 recycling, and crustal assimilation. *J. Volcanol. Geotherm. Res.* 206, 136–147.
445
- 446 Gelman, S. E., Gutierrez, F. J., Bachmann, O., 2013. On the longevity of large upper crustal silicic
447 magma reservoirs. *Geology* 41, 759-762.
448
- 449 Ghiorso, M.S., Gualda, G.A., 2013. A method for estimating the activity of titania in magmatic
450 liquids from the compositions of coexisting rhombohedral and cubic iron–titanium
451 oxides. *Cont. Min. Pet.* 165, 73-81.

- 452
453 Glazner, A.F., Coleman, D.S., Mills, R.D., 2015. The Volcanic-Plutonic Connection. In: Adv.
454 Volc. 1-22. Springer-Verlag
455
- 456 Green, T.H., Pearson, N.J., 1986. Ti-Rich Accessory Phase Saturation In Hydrous Mafic-Felsic
457 Compositions at High P-T. Chem. Geol. 54, 185-201.
458
- 459 Hildreth, E.W., 1981. Gradients in silicic magma chambers: Implications for lithospheric
460 magmatism. J. Geophys. Res. 86, 10,153–10,192.
461
- 462 Hildreth, W., 2004. Volcanological perspectives on Long Valley, Mammoth Mountain, and Mono
463 Craters: several contiguous but discrete systems. J. Volc. Geotherm. Res. 136, 169–
464 198.
465
- 466 Huber, C., Bachmann, O., Manga, M., 2012. Two competing effects and volatiles on heat
467 transfer in crystal-rich magmas: thermal insulation vs defrosting. J. Pet. 51, 847-867.
468
- 469 Kennedy, B., Wilcock, J., Stix, J., 2012. Caldera resurgence during magma replenishment and
470 rejuvenation at Valles and Lake City calderas. Bull. Volc. 74, 1833–1847.
471
- 472 Kern J.M., de Silva, S.L., Schmitt, A.K., Kaiser, J.F., Iriarte, A.R., and Economos, R., 2016.
473 Geochronological imaging of an episodically constructed subvolcanic batholith” U-Pb in
474 zircon chronochemistry of the Altiplano Puna Volcanic Complex of the Central Andes.
475 Geosphere, v.12, no.4, 24pp.
476
- 477 Lipman, P.W., 2007. Incremental assembly and prolonged consolidation of Cordilleran magma
478 chambers: Evidence from the Southern Rocky Mountain volcanic field. Geosphere 3, 42-
479 70.
480
- 481 Lindsay, J. M., Schmitt, A. K., Trumbull, R. B., de Silva, S. L., Siebel, W., Emmermann, R., 2001.
482 Magmatic evolution of the La Pacana caldera system, Central Andes, Chile:
483 Compositional variation of two cogenetic, large-volume felsic ignimbrites. J. Pet. 42,
484 459–486.
485
- 486 Liu, Y., Hu, Z., Zong, K., Gao, C., Gao, S., Xu, J., Chen, H., 2010. Reappraisal and
487 refinement of zircon U-Pb isotope and trace element analyses by LA-ICP-MS. Chinese
488 Sci. Bull. 55, 1535-1546.
489
- 490 Linnen, R. L., & Keppler, H., 2002. Melt composition control of Zr/Hf fractionation in magmatic
491 processes. Geochim. Cosmochim. Acta, 66(18), 3293-3301.
492
- 493 Ludwig, K.R., 2008. Isoplot/Ex version 3.75, a geochronological toolkit for Microsoft Excel.
494 Berkeley Geochronology Center Special Publication.

- Luhr, J.F. and Carmichael, I.S.E., 1980. The Colima volcanic complex, Mexico. I: post-caldera andesites from Volcan Colima. *Cont. Min. Pet.* 71: 343-372.
- Mahon, K. I., 1996. The New "York" regression: Application of an improved statistical method to geochemistry. *Int. Geol. Rev.* 38, 293-303.
- Mahood, G.A. and Hildreth, E.W., 1983. Large partition coefficients for trace elements in high-silica rhyolites. *Geochim. Cosmochim. Acta* 47: 11-30. doi: 10.1016/0016-7037(83)90087-X.
- Marsh, B.D., 1984. On the Mechanics of Caldera Resurgence. *J. Geophys. Res.* 89, 8245-8251.
- Paces, J. B., Miller, J.D., 1993. Precise U-Pb ages of Duluth complex and related mafic intrusions, northeastern Minnesota: Geochronological insights to physical, petrogenetic, paleomagnetic, and tectonomagmatic processes associated with the 1.1 Ga midcontinent rift system. *J. Geophys. Res.* 98, 13997-14013.
- Pamukcu A.S., Gualda G.A.R., Bégué F., Gravley D.M., 2015. Melt inclusion shapes: Timekeepers of short-lived giant magma bodies. *Geology* 43, 947-950
- Pearce, N. J., Perkins, W. T., Westgate, J. A., Gorton, M. P., Jackson, S. E., Neal, C. R., Chenery, S. P., 1997. A compilation of new and published major and trace element data for NIST SRM 610 and NIST SRM 612 glass reference materials. *Geostand. Newslett.* 21, 115-144.
- Reid, M.R., 2003. Timescales of Magma Transfer and Storage in the Crust. *Treatise Geochem.* 3, 167-193.
- Reid, M.R., Coath, C.D., Harrison, T.M. McKeegan, K.D., 1997. Prolonged residence times for the youngest rhyolites associated with Long Valley Caldera: ^{230}Th - ^{238}U ion microprobe dating of young zircons. *EPSL.* 150, 27-39.
- Reid, M.R., Vazquez, J.A., Schmitt, A.K., 2011. Zircon-scale insights into the history of a Supervolcano, Bishop Tuff, Long Valley, California, with implications for the Ti-in-zircon geothermometer. *Cont. Min. Pet.* 161, 293-311.
- Salisbury, M.J., Jicha, B.R., de Silva, S.L., Singer, B.S., Jimenez, N.C., Ort, M.H., 2011. $^{40}\text{Ar}/^{39}\text{Ar}$ chronostratigraphy of Altiplano-Puna volcanic complex ignimbrites reveals the development of a major magmatic province. *Geol. Soc. Am. Bull.* 123, 821-840.
- Sambridge, M.S., Compston, W., 1994. Mixture modeling of multi-component datasets with application to ion probe zircon ages. *EPSL.* 128, 373-390.

- Sano, Y., Terada, K., & Fukuoka, T., 2002. High mass resolution ion microprobe analysis of rare earth elements in silicate glass, apatite and zircon: lack of matrix dependency. *Chem. Geol.*, 184(3), 217-230.
- Schmitt, A.K., Lindsay, J.M., de Silva, S., Trumbull, R.B., 2002. U-Pb zircon chronostratigraphy of early-Pliocene ignimbrites from La Pacana, north Chile: implications for the formation of stratified magma chambers. *J. Volc. Geotherm. Res.* 120, 43-53.
- Smith, R.L., Bailey, R.A., 1968. Resurgent Cauldrons. *Geol. Soc. Am. Mem.* 116, 613- 662.
- Smith, R.L., 1979. Ash-flow magmatism. *Geol. Soc. Am. Spec. Pap.* 180, 5-27.
- Ward, K. M., Zandt, G., Beck, S. L., Christensen, D. H., McFarlin, H., 2014. Seismic imaging of the magmatic underpinnings beneath the Altiplano-Puna volcanic complex from the joint inversion of surface wave dispersion and receiver functions. *EPSL.* 404, 43–53.
- Watts, R.B., de Silva, S.L., Jimenez de Rios, G., Croudace, I., 1999. Effusive eruption of viscous silicic magma triggered and driven by recharge: a case study of the Cerro Chascon-Runtu Jarita Dome Complex in Southwest Bolivia. *Bull. Volc.* 60, 241-264.
- Wotzlaw, J. F., Schaltegger, U., Frick, D. A., Dungan, M. A., Gerdes, A., Günther, D., 2013. Tracking the evolution of large-volume silicic magma reservoirs from assembly to supereruption. *Geology* 41, 867-870.
- Wotzlaw, J. F., Bindeman, I. N., Watts, K. E., Schmitt, A. K., Carrichi, L., Schaltegger, U., 2014. Linking rapid magma reservoir assembly and eruption trigger mechanisms at evolved Yellowstone-type supervolcanoes. *Geology* 42, 807–810.

FIGURE CAPTIONS

Figure 1: Geologic Map of the Pastos Grandes Caldera. N, M, and S, refer to the Northern, Middle, and Southern Post-Climactic Domes.

Figure 2: Classification Diagram. Symbols include all volcanic and plutonic rocks of the caldera complex. The majority are classified as high K dacites, while fewer samples are classified as high K andesites and rhyolites. All samples related to the PGI cycle, including the post-climactic domes and xenoliths fall into the category of high K dacite (64-68 wt% SiO₂). The more evolved

xenoliths are related to the Chuhuilla stage of activity. Representative volcanic and plutonic samples of the pre- and post-climactic magma stages are included as well. Increasing crystallinity is seen from the PGI through the Post-Climactic Domes, to the Granodiorite xenoliths.

Figure 3: Representative cathodoluminescence images of zircon crystals from each of the samples. A & B are from the Pastos Grandes Ignimbrite. Whereas most analyses represent rim ages, core/rim pairs were compared in selected crystals. C, D, & E are from the South, Middle and North Post-Climactic Dome respectively. F & G from the plutonic xenolith. H&I illustrate the xenocrystic cores that make up roughly 2% of the total population.

Figure 4: (A) $^{206}\text{Pb}/^{238}\text{U}$ zircon ages for pre- and post-climactic Pastos Grandes samples shown as individual points in ranked order, a histogram, and a probability density curve. Pre-climactic samples comprise PGI and South PCD, post-climactic samples Mid and North PCD as well as granodiorite. The two major peaks make up the magmatic stages that are separated in the lower panel (B). (B) U-Pb zircon age probability density curves and a ranked order plot for PGI, three Post-Climactic Domes, and granodiorite clasts zircons ($n = 186$). The PGI (peak age: 3.19 ± 0.04 Ma) and the South Post-Climactic Dome (peak age: 3.11 ± 0.03 Ma) make up the pre-climactic magma stage. The Middle (peak age: 2.51 ± 0.04 Ma) and Northern (2.54 ± 0.05 Ma) Post-Climactic Domes as well as the granodiorite (2.63 ± 0.01 Ma) contain similar zircon age populations in the post climactic magma stage. All errors are 1 sigma.

Figure 5: Zircon trace element variation diagrams for pre- and post- climactic Pastos Grandes samples. A) Ti vs. Zr/Hf correlate consistent with zircon fractional crystallization and cooling. Because pre- and post- climactic samples overlap over the same temperature range and display identical fractionation trend, a secular cooling trend for the Pastos Grandes magma system can be ruled out. B) Eu/Eu* vs. Zr/Hf. C) U/Th vs. Zr/Hf. D) Yb/Gd vs. Zr/Hf. Inset shows schematic fractional crystallization trends drawn to scale with the main panels involving major (pl = plagioclase, hbl = hornblende, qtz = quartz, and bt = biotite) and accessory minerals (zrn = zircon, sph = sphene, all = allanite). Increasing U/Th and Yb/Gd with decreasing Zr/Hf (equivalent to decreasing temperature) suggests that Pastos Grandes magmas evolved while accessory minerals (sph, all) were stable. Major mineral abundances were: pl = 45%, hbl = 35%, qtz = 10%, bt = 5%; zircon was included at zrn = 0.1%. Trends with accessory minerals present were computed at the following abundances: sph = 5%, all = 0.3% (all percentages by mass).

Figure 6: Zircon compositions (A = Zr/Hf; B = Eu/Eu*; C = U/Th; D = Yb/Gd) plotted against $^{206}\text{Pb}/^{238}\text{U}$ age. Two-dimensional probability density contours are shown with warm colors indicating higher probabilities. Zircon compositional parameters are temperature sensitive, with schematic cooling trends indicated by arrow (i.e., Zr/Hf and Eu/Eu* are expected to decrease with falling magmatic temperatures, whereas U/Th and Yb/Gd tend to increase with cooling due to accessory sphene and allanite becoming stable). Although pre-and post-climactic zircons overlap in all diagrams, there is a slight shift indicating conditions of zircon crystallization at higher temperatures for the post-climactic zircons. This is interpreted to indicate a reheating event following the climactic eruption of the PGI (dashed vertical line).

618

619 Figure 7: Ti-in zircon model temperatures (at aTiO₂ = 0.7) vs. ²⁰⁶Pb/²³⁸U age. Two-dimensional
620 probability density contours are underlain (see caption to Figure 6). PGI climactic eruption age
621 is indicated by dashed vertical line. Ti-in-zircon temperatures overlap between pre- and post-
622 climactic zircon (see Fig. 5), but there is also a significant proportion of post-climactic zircon
623 that crystallized at elevated temperatures (by 50-100°C) compared to the highest-probability
624 cluster in the pre-climactic zircon population. This suggests that at least some post-caldera
625 zircons crystallized from an initially hotter magma, following re-establishment of magmatic
626 equilibrium by ascent of deeper and hotter magma after the PGI caldera collapse.

627

628

629 Table 1: Summary of Zircon Chronochemistry.

630

631

Figure

Figure

Figure

Figure

Figure

Increasing Textural Maturity

PGI Pumice (~40% Crystals)

PCD (45-50% Crystals)

CCG (Holocrystalline)

Figure

Figure

Table

Sample/ Analysis	206Pb/ 238U Age (Ma)	±1σ Ma	238U/ 206Pb	±1σ	207Pb/ 206Pb	±1σ	% 206Pb	U	Th	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb
PGI 1	3.29	0.14	2017	90.35	0.05	0.003	99.90	2244	624	0.09	14	0.09	0.20	1.91	1.99	10.6	26.7	52	108	198	367	648
PGI 2*	3.76	0.18	1729	81.87	0.06	0.006	98.26	1893	245	0.80	279	18.6	55.2	241	174	763	1474	2460	4000	6041	8603	11189
PGI 3	3.03	0.12	2157	82.82	0.06	0.004	98.70	2075	588	-	-	-	-	-	-	-	-	-	-	-	-	-
PGI 4	3.15	0.13	1871	70.71	0.13	0.007	88.89	2511	1158	1.23	35	1.04	1.26	6.35	6.03	28.9	26.7	139	284	542	945	1602
PGI 5	3.16	0.11	2064	73.70	0.06	0.004	98.02	2377	446	0.10	26	0.22	0.80	6.10	5.01	29.3	66.6	130	264	459	790	1236
PGI 6	4.10	0.18	1594	70.60	0.05	0.003	99.01	2213	578	1.12	33	1.59	3.42	14.1	9.89	61.1	124.1	219	408	691	1111	1680
PGI 7	3.10	0.14	2085	89.99	0.07	0.008	97.16	1839	482	0.54	44	0.81	0.71	6.13	4.88	27.1	61.6	126	268	487	863	1398
PGI 8	3.03	0.13	2163	93.11	0.06	0.006	98.56	2165	510	0.15	47	0.63	1.91	12.0	8.63	67.4	146.7	266	488	823	1331	1974
PGI 9	7.55	0.32	855	36.52	0.05	0.003	99.03	2422	686	0.82	37	1.26	4.39	27.3	14.6	119	268	505	903	1456	2189	3008
PGI 10	3.19	0.14	2061	89.20	0.05	0.005	98.88	2614	783	-	-	-	-	-	-	-	-	-	-	-	-	-
PGI 11	3.16	0.26	1189	49.66	0.39	0.010	56.58	1505	347	0.68	77	1.38	3.10	20.7	13.4	82.4	184.8	338	615	1022	1635	2292
PGI 13	3.08	0.17	1780	81.40	0.18	0.006	82.46	2385	754	0.19	46	0.37	1.11	9.71	5.20	36.9	86.7	185	376	690	1195	1841
PGI 14	3.41	0.14	1948	80.81	0.05	0.004	100.13	2335	580	0.22	36	0.38	1.20	7.08	5.74	32.2	71.0	137	273	466	791	1271
PGI 15	3.59	0.16	1816	78.84	0.06	0.004	98.29	1893	383	2.18	30	1.38	2.41	7.11	5.77	35.4	79.2	147	278	456	738	1052
PGI 16	3.35	0.15	1948	86.54	0.06	0.003	98.37	2622	921	0.95	32	0.71	1.44	9.25	5.49	36.8	85.3	165	315	568	945	1515
PGI 17	3.07	0.75	444	17.67	0.67	0.011	20.46	1063	243	0.07	30	0.35	0.86	5.93	4.60	24.0	57.2	105	209	371	633	1034
PGI 18	3.53	0.13	1873	70.16	0.05	0.004	99.89	2078	616	0.40	358	4.60	15.5	83.5	55.7	299	587	972	1551	2275	3218	4105
PGI 19	3.22	0.14	2037	89.25	0.06	0.004	98.71	2330	670	-	-	-	-	-	-	-	-	-	-	-	-	-
PGI 20	3.52	0.15	1844	79.23	0.06	0.004	98.01	2009	613	0.12	37	0.34	0.79	6.73	4.49	26.5	56.0	111	205	376	632	1059
PGI 21	3.20	0.14	2044	90.22	0.06	0.004	98.54	2689	892	1.10	89	1.38	2.11	7.87	7.02	32.6	80.8	154	321	602	1081	1751
PGI 22	3.52	0.16	1856	83.33	0.06	0.003	98.81	2101	772	-	-	-	-	-	-	-	-	-	-	-	-	-
PGI 23	3.22	0.15	2025	56.16	0.06	0.004	93.94	2280	1168	0.16	39	0.52	1.47	8.94	6.18	33.2	70.5	133	259	453	760	1151
PGI 24	3.05	0.06	2106	41.65	0.07	0.005	96.68	1545	469	0.17	45	0.68	1.28	9.32	5.39	38.4	77.2	146	271	474	783	1204
PGI 25	3.57	0.10	1814	52.33	0.06	0.004	97.78	2286	404	0.19	36	0.43	1.20	5.75	4.43	31.2	63.1	120	234	415	740	1120
PGI 26	3.71	0.11	1733	50.48	0.07	0.007	97.19	1182	197	0.31	35	0.61	1.10	5.73	5.99	26.2	59.2	105	205	367	634	1023
PGI 27	4.61	0.16	1412	49.82	0.05	0.004	99.00	1937	670	-	-	-	-	-	-	-	-	-	-	-	-	-
PGI 28	3.91	0.10	1667	43.63	0.06	0.004	98.69	2135	471	0.19	38	0.54	1.02	6.98	5.45	30.9	70.2	142	282	535	917	1464
PGI 29	3.00	0.10	2135	70.68	0.08	0.008	96.05	1326	282	1.03	31	0.88	1.70	4.43	4.65	23.1	49.5	97	195	355	634	995
PGI 30	3.61	0.08	1797	40.05	0.06	0.003	98.47	4420	2583	0.21	32	0.52	0.78	6.73	4.65	26.0	55.1	104	202	375	654	1044
PGI 31	4.17	0.19	1336	50.32	0.17	0.007	84.50	934	173	-	-	-	-	-	-	-	-	-	-	-	-	-
PGI 32	3.63	0.07	1786	36.05	0.06	0.003	98.04	3298	992	0.22	54	0.71	1.55	10.1	7.63	45.2	96.3	168	311	504	791	1219
PGI 33	3.19	0.08	2025	49.62	0.07	0.005	97.22	1968	403	0.39	59	0.79	2.20	12.5	8.86	52.2	109.0	193	359	602	957	1423