Geomorphic expression of rapid Holocene silicic magma reservoir growth beneath Laguna del Maule, Chile

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Large rhyolitic volcanoes pose a hazard, yet the processes and signals forecasting an eruption are obscure. Satellite geodesy has revealed surface inflation signaling unrest within magma reservoirs underlying a few rhyolitic volcanoes. Although seismic, electrical, and potential field methods may illuminate the current configuration and state of these reservoirs, they cannot fully address the processes by which they grow and evolve on geologic time scales. We combine measurement of a deformed paleoshore surface, isotopic dating of volcanic and surface exposure, and modeling to determine the rate of growth of a rhyolite-producing magma reservoir. The numerical approach builds on a magma intrusion model developed to explain the current, decade-long, surface inflation at >20 cm/year. Assuming that the observed 62-m uplift reflects several non-eruptive intrusions of magma, each similar to the unrest over the past decade, we find that ~13 km$^3$ of magma charged the reservoir at a depth of ~7 km during the Holocene, accompanied by the eruption of ~9 km$^3$ of rhyolite. The long-term rate of magma input is consistent with reservoir freezing and platoon formation. Yet, the unique set of observations considered here implies that large reservoirs can be incubated and grow at shallow depth via episodic high-flux magma injections. These replenishment episodes likely drive rapid inflation, destabilize cooling systems, propel rhyolitic eruptions, and thus should be carefully monitored.

INTRODUCTION

The gradual accumulation of rhyolitic magma in the uppermost crust can promote large caldera-forming eruptions (1, 2), but the processes by which this occurs remain poorly understood (3, 4). A key issue is whether the magma flux into the upper crust is sufficiently large, over long enough periods of time, to sustain growth of magma reservoirs thermally capable of producing large rhyolitic eruptions, rather than crystallizing into plutons (5–8). Many of the observations fueling this debate come from radioisotopic dating of minerals [for example, (8)] or trace element diffusion clocks preserved within them [for example, (4)]. On the other hand, magma fluxes typical of plutonic systems are thought to normally be too low to sustain large, eruptible magma reservoirs in the upper crust (6, 9–12). On the other hand, a growing body of evidence suggests that silicic magma is stored long-term at relatively cool, nearly subsolidus conditions and is episodically remobilized by rapid injection of hot recharge magma that may propel destabilization and eruption (3, 4, 8, 13–15).

The magnitude and pattern of surface deformation offer another important accumulation of magmatic processes operating beneath restless, occasionally active, volcanoes (16, 17). Measurement of deformation by satellite geodesy in caldera volcanoes that produced supereruption scale (18, 19), as well as modest volume (20–23) rhyolitic eruptions, has revealed inflation affecting regions of hundreds of square kilometers, over periods of months to years, typically at rates of 10 cm/year or less (17). Arrival of new magma into an extent, shallow magma reservoir is the common explanation for the surface inflation [for example, (22)], although the pressurization of magmatic fluids also contributes in some systems (17, 24). What has remained elusive is a means of measuring a long-term flux rate that integrates several magma pulses and that may leverage interpretations of whether a magma reservoir is likely to grow and erupt or to freeze into a platoon (7).

The Laguna del Maule (LdM) volcanic field (Fig. 1) comprises the greatest concentration of postglacial (younger than ~20,000 years) rhyolite in the Andes and includes the products of ~40 km$^3$ of explosive and effusive eruptions (25–28). Recent observations at LdM by interferometric synthetic aperture radar (InSAR) and global positioning system (GPS) satellite geodesy have revealed inflation at rates exceeding 20 cm/year since 2007 (29–31), thereby capturing an ongoing period of growth of a potentially large upper crustal magma reservoir (27). The current episode of inflation has been explained by a model of transient supply of magma into this reservoir at a depth of 4.5 km and requiring recharge at a rate of 0.03 to 0.04 km$^3$/year (31, 32), which is a flux sufficient to destabilize a cool silicic magma reservoir (6, 7, 14, 22). Here, we use a geomorphic record of surface deformation at LdM that offers an unprecedented opportunity to link this current episode of unrest and inflation in a rhyolite-producing system to the record of rhyolitic volcanism and magma intrusion spanning the last 10,000 years.

RHYOLITIC VOLCANISM AND GEOMORPHIC EVOLUTION

Rhyolitic volcanism and unrest

LdM sits atop the southern Andean range crest where, following a rapid retreat of glaciers between ~23 and 19 thousand years ago (based on $^{40}$Ar/$^{39}$Ar dating of pre- and postglacial lava flows), a
flare-up of dominantly silicic volcanism occurred within the central lake basin (Fig. 1) (25, 33, 34). The silicic eruptions include effusive and explosive events and are volumetrically dominated by crystal-poor rhyolite. Rhyodacite and andesite eruptions also occurred throughout postglacial time but are concentrated in the northwestern to western LdM basin several kilometers from the locus of rhyolitic volcanism (Fig. 1). Rhyodacitic lavas are distinguished from rhyolitic ones by higher crystallinities, common amphibole crystals, and abundant centimeter-scale chilled mafic inclusions, which are nearly absent within the rhyolites (25, 33, 34). 

The presumed vent for the plinian eruption of the Rhyolite of LdM (unit rdm) is beneath the lake. The star in the inset shows the location of LdM in South America. Important map units include (i) rle, Rhyolite of Los Espejos that damned the lake at 19 thousand years; and (ii) rcb, the many rhyolitic lavas of the Barrancas complex that were emplaced during the last 14 thousand years. These and other map unit abbreviations follow those in (25, 34).

Each cubic kilometer of erupted rhyolite is distilled from ~3 to 4 km³ previously emplaced magma ± juvenile crust at depths of 4 to 6 km (33, 34). Amphibole barometry, trace element compositions, thermodynamic modeling of phase equilibria, and Sr, Pb, and Th isotope ratios indicate that crystal-poor rhyolite erupted at LdM was produced in an outbreak flood of 15 km³ of water pouring through the gorge and scouring the upper Maule river valley (25). The lava dam was breached, resulting in an outbreak flood of 15 km³ of water pouring through the gorge and scouring the upper Maule river valley (25). The highstand shoreline is variably preserved and expressed around the basin. Atop the Espejos rhyolite coulée, it is a beach deposit of rounded boulders of the rhyolite (Fig. 3A). To the northeast, it is a bench up to several tens of
meters wide cut directly into Pleistocene volcanic rocks (Fig. 3B). Around the eastern and southern edges of the basin, it is a bench cut into older Pleistocene lavas, covered by rounded beach boulders and late Holocene tephra (Fig. 3C). To determine the age of the highstand shoreline, we collected samples for surface exposure dating from five wave-cut terrace outcrops in the northern LdM basin at elevations between 2374 and 2388 meters above sea level (masl) (Fig. 2). Cosmogenic $^{36}$Cl measurements (Materials and Methods) yield ages of 9.4 ± 0.4, 8.8 ± 0.6, 7.5 ± 0.3, 6.6 ± 0.6, and 4.2 ± 0.2 ka (±2σ analytical uncertainties; table S1A). Two outcrops, one in rhyodacite ignimbrite at 2376 masl (Fig. 3B), the other 4 km to the northeast in andesite at 2386 masl, comprise tens to hundreds of square meters of block-jointed rock in windswept locations and yield the oldest dates of 9.4 ± 0.3 ka and 8.8 ± 0.6 ka that are indistinguishable from one another at the 95% confidence level. The three other sites on the rhyodacite ignimbrite spanning the same range in elevation yield younger ages. For samples that contain low native chlorine (table S1C), erosion of the surface or burial under ash, snow, or water will lower apparent $^{36}$Cl surface exposure ages, whereas for samples with high Cl, the impacts of erosion and burial tend to increase the apparent exposure ages. We therefore infer that the two younger ages from the shoreline derived from samples with low Cl (7.5 ± 0.3 thousand years ago and 4.2 ± 0.2 thousand years ago) reflect erosion or burial and that each gives a minimum duration of exposure. Likewise, the oldest $^{36}$Cl exposure

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**Fig. 2.** Map of preserved highstand paleoshoreline (color-contoured for elevation) and postglacial silicic lava flows. Map unit abbreviations follow those in (25, 33, 34). ka, thousand years ago (ka ago).
age from the shoreline (9.4 ± 0.3 thousand years ago), also derived from a low-Cl sample, provides a minimum age of exposure. However, we consider this figure to be a closely limiting minimum age because posited rock surface erosion rates have a negligible impact (table S1A), and past cover by ash or snow is deemed minimal because of the windswept location of these outcrops. The exposure ages derived from moderate- to high-Cl material (8.8 ± 0.6 thousand years ago and 6.6 ± 0.6 thousand years ago) are more difficult to evaluate with regard to erosion, past cover, and pore water content (table S1A) but may in part reflect episodic submergence associated with sea-level changes similar to that observed along the shore of the modern lake (25).

We conclude that the abrupt drop in lake level to near its modern elevation occurred at ca. 9.4 thousand years ago and left the highstand shoreline outcrops exposed to buildup of 36Cl.

Measuring Holocene surface deformation
We measured GPS positions of 64 sites on the highstand surface in a rapid static mode using five continuously recording GPS stations as base stations (Fig. 2), yielding elevations with centimeter-scale precision (Materials and Methods and table S2). We corroborated these GPS measurements by mapping each onto a photogrammetrically generated 1-m-resolution digital elevation model (DEM; Materials and Methods) and outlining the preserved highstand shoreline outcrops (Fig. 2 and fig. S1). Most surfaces are inclined toward the modern lake, and many are covered by tephra (Fig. 3C); hence, the main uncertainty in the GPS measurements comes from the choice of the GPS site location within one inclined “outcrop,” which may extend over 50 to 100 m in a horizontal distance. The comparison between the DEM and GPS height measurements allows us to estimate a total uncertainty for each GPS position of ±3 m (Materials and Methods and fig. S2). Elevations of the 64 GPS sites show 62 m of relief and range from 2366 to 2428 masl, with the lowest values along the western and northern perimeter of the lake basin and the highest along the south and east (Fig. 4 and fig. S1A).

INTERPRETATION AND MODELING OF GROUND DEFORMATION
The highstand paleoshoreline is inferred to have been a horizontal surface when the lake drained at 9.4 thousand years ago. Its elevation gradient and spatial distribution are inconsistent with predicted isostatic rebound in response to draining of the lake [for example, (39)] (Materials and Methods and fig. S3). The LdM basin is adjacent to the Andean range crest with current elevations of 2100 to 3000 masl throughout the region shown in Fig. 1. Here, during the Last Glacial Maximum ~21 thousand years ago, an ice cap ~80 km across and up to several hundred meters thick occupied the deepest valleys and range crest basins (40, 41). Deglaciation of the LdM basin was likely completed by 19 to 18 thousand year ago, resulting in isostatic adjustment to less than 5 MPa (megapascals) equivalent of lost ice. Modeling of glacial isostatic
uplift of the European Alps (42), where the extent and thickness of ice at the Last Glacial Maximum were much larger than that at LdM, indicates that deglaciation induced about 2 mm/year of uplift. Applying this estimate as a maximum for LdM implies that less than 20 m of uplift, across the entire region shown in Fig. 1, may have occurred during the last 9.4 thousand years (Materials and Methods). Instead, the gradient is consistent with deformation associated with an inflating magma reservoir located below the southeastern portion of the lake basin near the Barrancas volcanic complex (Fig. 2). Moreover, the 60+ m of permanent deformation during the last 9.4 thousand years favors intrusion and solidification of magma, rather than pressurization of volatiles in a hydrothermal system.

To investigate the origin of the deformed paleoshoreline, we model the cumulative vertical displacement measured over the last 9.4 thousand years and relate it to the numerical modeling results of magma injection into a long-lived reservoir on the decadal time scale (32). The final GPS elevation data set is referenced in space to the lowest elevation in the northwest (Fig. 4) and inverted for three different deformation source geometries. Assuming that the crust is isotropic, homogeneous, elastic half-space, we find the best-fitting source to be a prolate spheroid located in the southeast of the lake basin experiencing a volume change of 13 km$^3$ over 9.4 thousand years at a depth of 7 km (Figs. 2 and 4A). A statistical F test confirms that this model with seven adjustable parameters is a significantly better fit than a simple spherical source with 95% confidence (Materials and Methods). The inflating source model reproduces the main characteristics of the spatial pattern and magnitude of the paleoshoreline deformation. The location of the inferred source of magma intrusion is centered among the rhyolitic volcanic complexes that ring the southern and eastern...
flanks of the LdM basin and from which eruptions were focused during the last 9.4 ka (Figs. 1 and 2).

Whereas there are no obvious offsets of the highstand paleoshoreline bench by faults, there is a relatively abrupt change in elevation of 20 m over a distance of 500 m across the inferred trace of the regional Troncoso fault (Fig. 2 and fig. S4). Cumulative movement of several meters along this fault may explain some of the larger residuals visible on the radial profile (Fig. 4C).

The kinematic model considers a constant change in pressure with time to reproduce the pattern of cumulative uplift. In the absence of a record of the temporal evolution of the uplift over the 9.4 thousand-years time interval, we use information revealed from the study of the ongoing uplift episode that began between 2004 and 2007. To reproduce the exponentially increasing and decreasing rates of uplift measured at LdM, Le Mével et al. (32) modeled the flow of magma through a conduit into an ellipsoidal source underlying most of the central part of the lake basin. If several such non-eruptive deformation episodes with the same characteristic time constant occurred over 9.4 thousand years ago, then 16 distinct episodes such as this would be necessary to uplift the paleoshoreline by 60 m as plotted in Fig. 5. However, the modeling results indicate two main differences between the deformation sources: The location has moved northwestward since 9.4 thousand years ago (Fig. 2), and the Holocene deformation source was deeper, leading to a wider deformation footprint at the surface.

**DISCUSSION**

**Geomorphologic constraints on magma reservoirs**

Geomorphology has been used to constrain the depths and growth rates of magma reservoirs in a variety of settings, a few examples of which are outlined here. Whereas InSAR data show that growth of the magma body at a depth of 19 km below Socorro, New Mexico,
props recent uplift at 2.5 mm/year, geomorphic evidence suggests that it is unlikely that this uplift has persisted for more than a few centuries (43). Geodetically measured uplift rates at Uturuncu and Lazeuf volcanoes in the Central Andes are between 1.0 and 3.5 cm/year (44). Several lines of geomorphic evidence suggest that at Lazeuf, expansion of a magma reservoir at depths of 10 to 20 km generated pulses of surface uplift that total several hundred meters during the last 400 thousand years, punctuated by long intervals of quiescence or subsidence (45). Also in the Central Andes, a long-wavelength topographic dome about 1 km high reflects growth of the enormous 10- to 20-km-deep Altiplano-Puna magma body during perhaps the last 10 million years (Ma) (46). Resurgent doming or uplift of tens to hundreds of meters over periods of centuries to several millennia at rates of between 1 and 20 cm/year has also been ascribed to magma intrusion beneath calderas at Toba (47), Campi Flegrei (48), and Iwo Jima (49). Features that distinguish LdM from these other examples include the following: (i) the shallow depth of the magma reservoir inferred from gravity (36), geodesy (30–32), and petrology (33, 34); (ii) a highly resolved series of rhyolitic eruptions of well-constrained volumes before, and during, the 60+ m of surface deformation of the last 9.4 thousand years (Fig. 1); and (iii) the exceptional magnitude of the current uplift at more than 20 cm/year during the last decade that is centered among the many rhyolitic vents (Fig. 2). We next integrate these observations to explore the long-term dynamics of the LdM magma reservoir.

**Incremental growth and eruption of a shallow plutonic reservoir**

Deformation of the paleoshoreline, coupled with knowledge of the volcanic output, the numerical simulations of magma intrusion, and geometry of the Bouguer gravity anomaly (33), leads to the model outlined in Fig. 6. During the last 9.4 thousand years, the ratio of rhyolite erupted (~9 km$^3$) to the volume of magma transiting into and out of the reservoir (22 km$^3$) is about 0.4, suggesting reservoir growth through underaccretion of magma as suggested in some episodal plutons (6, 50, 51). Thermal modeling suggests that at a depth of ~7 km, the time-averaged magma flux here of 0.0023 km$^3$/year is unlikely to favor the eruption of rhyolite, instead leading to solidification of a pluton (7, 10, 52). However, the >1.5 million years of LdM volcanism (25) attests to a magma flux that has thermally primed the upper crust to host a large, dynamic, eruptible reservoir of silicic magma [for example, (11, 33, 34)]. During the Holocene alone, at least 10 eruptions of crystal-poor rhyolite reflect crystallization of perhaps 40 km$^3$ of andesitic to rhyodacitic magma within the shallow reservoir (34).

The absence of measured deformation before 2007 (29, 30) demonstrates that the current inflation event reflects an episodic process operating beneath LdM. Following the uniformity principle, the long-term deformation measured here would also reflect successive uplift episodes. We propose that the magma supply is best understood as an aggregation of repeated high-flux episodes exemplified by the ongoing inflation. During one such episode, the transient uplift rate could be as high as 200 mm/year, corresponding to a magma recharge rate of as high as 0.04 km$^3$/year (Fig. 5) (32). These recharge episodes each propelled destabilization of the reservoir via the addition of volatiles and ascent of bubble-rich plumes of rhyolitic melt through low-porosity crystal mush (Fig. 6) (33, 53, 54). Moreover, the locus of magmatic intrusion, wavelength of surface inflation, and locations of rhyolitic eruptions have migrated several kilometers during the last 20 thousand years, highlighting the spatially heterogeneous, incremental growth of the magma reservoir. The Bouguer gravity anomaly implies that the reservoir today is partially molten under the lake (~1800 to 2400 kg/m$^3$), with much denser crystal mush or subsolidus material (~2700 kg/m$^3$) underlying the southern flank of the basin (36), suggesting that here it has cooled and begun to solidify (Fig. 6). Additional gravity measurements on the Barrancas complex and to its south are needed to test and refine this inference. The time-averaged rate of magma accumulation beneath LdM during the Holocene of 0.0023 km$^3$/year is equivalent to many well-dated plutons (7, 9, 10, 50, 52) but has likely been punctuated by numerous high-flux recharge events at rates 20 times faster. Our findings illustrate that the incremental growth of plutons comprising several hundred cubic kilometers, accompanied by eruption of tens of cubic kilometers of rhyolite, is not restricted to deep hot crust. Rather, plutons may grow in the uppermost crust via repeated high-flux additions of hot magma to a gradually expanding, mostly frozen, body of plutonic mush (4, 14, 34). Future explosive eruptions of modest to large volume are to be expected as a consequence of rapid recharge eventsakin to that observed today at LdM.

**MATERIALS AND METHODS**

**36 Cl dating**

For 36 Cl exposure age determinations, five samples were collected from horizontal surfaces of the lake highstand shoreline notched into 990 thousand-years-old welded rhyodacite tuff or 1.01-million-years-old andesite north of LdM, and four samples came from horizontal surfaces on rhyodacitic or rhyolitic lava flows that exhibit uneroded morphologically youthful flow structures (Figs. 2 and 3 and table S1A). Local topographic maxima were selected for sampling to minimize the effects of shielding, which was evaluated using an inclinometer. Samples were collected using a hammer and chisel at least 5 cm from adjacent vertical surfaces to avoid edge effects in calculating ages. Whole-rock separates, crushed to 125 to 250 μm, were prepared from the uppermost 1.75 to 3.0 cm of each sample and ultrasonically cleaned in deionized water. Cl was extracted in clean laboratory facilities at the University of New Hampshire following methods developed by Stone et al. (55) and modified by Lacciardi et al. (56). Crushed samples were pretreated with 2% HNO3 and spiked with an enriched 37 Cl tracer and then dissolved in HF-HNO3 solution. Upon complete digestion, insoluble fluoride compounds were removed by centrifugation, and Cl was precipitated as AgCl with the addition of AgNO3. The precipitate was further purified by redissolution in NH4OH and the addition of Ba(NO3)2 to precipitate sulfate as BaSO4. AgCl was then reprecipitated by the addition of 2 M HNO3 and AgNO3, washed repeatedly in deionized water, and dried.

The 35 Cl/37 Cl and 36 Cl/37 Cl ratios were measured at the Center for Accelerator Mass Spectrometry at the Lawrence Livermore National Laboratory (LLNL) facility. Procedural blanks contributed 0.1 to 5.0% error to the 36 Cl concentration errors in the unknowns, and appropriate blank corrections were made before age calculations. Ages were calculated using the online CRONUScale exposure age calculator and the LSDn scaling scheme (57–59). Sensitivity analyses were conducted using the CRONUScale calculator to isolate the potential impacts of rock surface erosion rates and pore water content on the apparent exposure ages (table S1A). A prescribed rock surface erosion rate of 5 mm per thousand years, which represents a plausible value determined on similar volcanic lithologies in the Quelccaya region of the Peruvian Andes (60), was seen to have a minimal effect on apparent exposure ages derived from samples with low native Cl [15-SLM-05, 22.9 parts per
million (ppm) Cl; SLM-16-28, 25.3 ppm; SLM-16-29, 26.2 ppm], including the sample that provides the closely limiting 9.4 thousand years minimum age of the highstand shoreline. Likewise, a prescribed 0.5 fraction of pore water content had negligible impact on ages derived from these low-Cl rocks. In contrast, the same prescribed surface erosion rate and pore water content were seen to have variable and significant impacts on the apparent exposure ages of surfaces derived from rocks with moderate to high Cl contents (table S1, A and C).

**Static GPS measurements**

GPS data that measure shoreline height were collected during two survey periods in 2015 and 2016 using two Trimble dual-frequency receivers: 5700 and NetR9 with Zephyr and Zephyr Geodetic antennas. Each GPS point was occupied for at least 10 min, following a rapid static method. To obtain precise coordinates for each point, we followed the GPS differential procedure (61), using, as reference, five continuous GPS (cGPS) stations operated by the Observatorio Volcanologico de los Andes del Sur (OVDAS) located around the lake [Fig. 2; base stations are MAU2, PUEL, LDMP, COLO, and NIE2 cGPS sites of (31)]. Precise coordinates of continuous stations were estimated using Trimble postprocessing service (Topcon) relative to the World Geodetic System 84 (WGS84) ellipsoid. These positions were reprocessed with Trimble Business Center and compared with results from long time series processed with GIPSY-OASIS software (release 6.3) from the Jet Propulsion Laboratory, yielding similar results. For each campaign, 2015 and 2016, this procedure ensured that ongoing surface deformation is accounted for. Each GPS point was processed by calculating baselines with respect to the nearest or the two nearest permanent-fixed stations (fig. S2). We processed baselines and obtained fixed solutions for 64 GPS points; when two baselines were processed for the same point, we adjusted both results for a robust solution. Results are shown in table S2. Most GPS sites were chosen at the midpoint of the inclined slope of the paleoshoreline surface or at the highest break in its slope. The mean difference between the elevations in the DEM and the GPS sites was 0.27 ± 1.07 m. Of the 64 GPS sites, 48 (75%) agreed within 1 m with the DEM. The minimum and maximum differences were 0.02 and 2.62 m, respectively. Thus, we attributed a total uncertainty for each GPS position of ±3 m (fig. S2).

**Digital elevation model**

A DEM was constructed using photogrammetric analysis of several thousand oblique, near-vertical, digital images acquired during two helicopter survey campaigns in 2017 and methods outlined in (62, 63). Ground control points used in the modeling include several purpose-placed markings and several of the seismic instruments deployed to study the LdM magma system. The DEM has a spatial resolution of 1 m.

**Isostatic adjustments to lake-level lowering and deglaciation**

To consider surface displacement due to the sudden drop in lake level at 9.4 thousand years ago, we applied an elastic formulation (64) to calculate the unloading due to a 200 m drop in the water table that followed the current shoreline. Even an extremely low value of crustal shear modulus (1 GPa) would only account for 14 m of uplift, limited to the lake basin with a maximum centered within the lake (fig. S3). Therefore, we conclude that this process does not explain the observed pattern of uplifted shoreline to the SE by 60+ m (Fig. 4A).

To calculate the amount of uplift caused by viscoelastic postglacial rebound, we considered a model for the European Alps (42). Their estimate of 2 mm/year in the Alps was likely much faster than what the same process would produce anywhere at LdM in the Andes. First, the Alps sustained a larger (800 × 250 km) and much thicker (~2 km) ice cap. Second, since the Andes comprise a magmatically active subduction zone, the lithosphere is likely hotter and thus less viscous than below the Alps. The latter effect would reduce the effective elastic thickness and thus the contribution to uplift by deglaciation relative to mantle flow (42). In the Alps, the observed glacial uplift spans spatial length scales of hundreds of kilometers (42). At LdM, however, the warping of the LdM highstand paleoshoreline occurs over a much shorter spatial length scale of the order of ~15 km.

Both of these unloading processes (melting glaciers or draining lake) would lead to a rate of uplift that is fairly constant over the time scale of a decade. Neither process can explain the uplift measured between 2007 and 2017 at LdM. As discussed previously, the rate of uplift between 2003 and 2004 is 0 ± 10 mm/year (29–31). We conclude that the doming of >60 m is instead driven by magma intrusion below the southern lake basin consistent with the eruptive history (Figs. 1 and 2).

**Analytical modeling and interpretation**

Before modeling, elevations were referred to a site in the NW assumed to be the original shoreline height. It is the furthest point from the source and has an elevation of 2372 m. The relative heights are the input for the analytical models. The horizontal displacement was known and was therefore set to 0 ± 30 m in the model input. All the models considered in this study assumed the crust to be an isotropic, homogeneous, elastic half-space. We tested three possible source geometries: (i) a sphere, implemented using the formulation of (65); (ii) a prolate spheroid (66); and (iii) a sill or penny-shaped crack using the three-dimensional Green function proposed by Fialko et al. (67). The inversion for the three different deformation sources was implemented in the dMODELS software package (68). The nonlinear inversion algorithm is a combination of a local optimization (interior-point method implemented in the MATLAB function fmincon) and a random search. For each set of parameters, 250 runs were calculated, and the parameter uncertainty estimates were then computed using a Jackknife resampling method.

We used analytical solutions of a simple deformation source in an elastic half-space to estimate the source volume change at depth from the cumulative surface uplift. Numerical models of the well-constrained temporal evolution of uplift for the ongoing episode of deformation started in 2007 helped us interpret the results in terms of magma volume estimates. Note that a magma injection model with a viscoelastic rheology for the crust yields up to 4% more displacement than the purely elastic model (32) and therefore would lead to a smaller estimate of magma injection volume for each episode. Considering the magma compressibility would, on the other hand, significantly affect the volume estimates of the magma injection. As shown in (32), the magma injection volume needed to explain the current uplift is twice as large if we assume a compressible magma (with $B = 2.1 \times 10^{-9} \text{ Pa}^{-1}$). Consequently, the incompressible magma volume estimates, including the 12.7 ± 0.9 km$^3$ prolate spheroid in Fig. 4B, are minimum estimates.

**SUPPLEMENTARY MATERIALS**

Supplementary material for this article is available at http://advances.sciencemag.org/cgi/content/full/4/6/eaat1513/DC1
REFERENCES AND NOTES

1. W. Hildreth, Volcanological perspectives on Long Valley, Mammoth Mountain, & Mono Craters: Several contiguous but discrete systems. J. Volcanol. Geotherm. Res. 136, 169–198 (2004).

2. C. J. N. Wilson, S. Blake, B. L. A. Charlier, A. N. Sutton, The 26.5 ka Orouni eruption, Taupo volcano, New Zealand: Development, characteristics and evacuation of a large rhyolitic magma body. J. Petro. 47, 35–69 (2006).

3. R. S. J. Sparks, K. Cashman, Dynamic magma systems: Implications for forecasting volcanic activity. Elements 13, 35–40 (2017).

4. A. E. Rubin, K. M. Cooper, C. B. Till, A. J. R. Kent, F. Costa, M. Bose, D. Gravley, C. Deering, J. Cole, Rapid cooling and cold storage in a silicic magma reservoir recorded in individual crystals. Science 356, 1154–1156 (2017).

5. S. E. Gelman, F. J. Gutierrez, O. Bachmann, On the longevity of large upper crustal silicic magma reservoirs. Geology 41, 759–762 (2013).

6. C. Annen, J. D. Bundy, J. Leuthold, R. S. Sparks, Construction and evolution of igneous bodies: Towards an integrated perspective of crustal magmatism. Lithos 230, 206–221 (2015).

7. T. Menand, C. Annen, M. de Saint Blanquat, Rates of magma transfer in the crust: Insights into magma reservoir recharging and pluton growth. Geology 43, 199–202 (2015).

8. D. Szymanszki, J.-F. Wotzlaw, B. S. Ellis, O. Bachmann, M. Guillong, A. von Quadt, Protracted near-solidus storage and pre-eruptive rejuvenation of large magma reservoirs. Nat. Geosci. 10, 777–782 (2017).

9. M. J. Tappa, D. S. Coleman, R. D. Mills, K. M. Samperton, The plutonic record of a silicic ignimbrite from the Late Tertiary volcanic field, New Mexico. Geochim. Geophy. Geosyst. 12, Q10011 (2011).

10. D. S. Coleman, R. D. Mills, M. J. Zimmerer, The pace of plutonism. Elements 12, 97–102 (2016).

11. O. Karakas, W. Degryseur, O. Bachmann, J. Dufek, Lifetime and size of shallow magma bodies controlled by crustal-scale magmatism. Nat. Geosci. 10, 446–450 (2017).

12. M. Barboni, B. Schoene, Short eruption window revealed by absolute crystal growth rates in a granitic magma. Nat. Geosci. 7, 524–528 (2014).

13. K. M. Cooper, A. J. R. Kent, Rapid remobilization of magmatic crystals kept in cold storage. Nature 506, 480–483 (2014).

14. T. H. Drutt, F. Costa, M. Dungan, B. Scaillet, Decadal to monthly timescales of magma transfer and reservoir growth at a caldera volcano, Nature 482, 77–82 (2012).

15. N. L. Andersen, B. R. Jicha, B. S. Singer, W. Hildreth, Incremental heating of Bishop Tuff beneath the active Laguna del Maule volcanic field, central Chile. J. Volcanol. Geotherm. Res. 58, 85–114 (2017).

16. N. L. Andersen, B. S. Singer, F. Costa, J. Fournelle, J. S. Herrin, G. N. Fabbro, Petrochronologic perspective on rhyolite unrest at Laguna del Maule, Chile. Earth Planet. Sci. Lett. 493, 57–70 (2018).

17. P. Srugoa, M. Elsiondo, M. Rosas, B. Singer, N. Andersen, Cerro Barrancas, Laguna del Maule volcanic field: Eruptive stratigraphy and hazard assessment, International Association of Volcanology and Chemistry of the Earth’s Interior General Assembly Abstracts (2017), p. 1043.

18. C. A. Miller, G. Williams-Jones, D. Fournire, J. Witter, 3D gravity inversion and thermodynamic modeling reveal properties of shallow silicic magma reservoir beneath Laguna del Maule, Chile. Earth Planet. Sci. Lett. 459, 14–27 (2017).

19. K. V. Cashman, R. J. S. Sparks, J. D. Bundy, Vertically extensive and unstable magmatic systems: A unified view of igneous processes. Science 355, eaag3055 (2017).

20. W. Hildreth, S. Moorbath, Crustal contributions to arc magmatism in the Andes of central Chile. Contrib. Mineral. Petrology 98, 455–489 (1988).

21. C. Y. Chen, A. C. Malof, Revisiting the deformed high shoreline of Lake Bonneville. Quat. Sci. Rev. 159, 169–189 (2017).

22. N. R. I. Hulton, R. S. Purves, R. D. McCulloch, D. E. Sugden, M. J. Bentley, The last glacial maximum and deglaciation in southern South America. Quat. Sci. Rev. 21, 233–241 (2002).

23. S. F. L. Watt, D. M. Pyle, T. A. Mather, The volcanic response to deglaciation: Evidence from glaciated arcs and a reassessment of global eruption records. Earth Sci. Rev. 122, 77–102 (2013).

24. J. Mey, D. Scherler, A. D. Wickert, D. L. Egholm, M. Tesauro, T. F. Schildgen, M. R. Strecke, Glacial isostatic uplift of the European Alps. Nat. Commun. 7, 13382 (2016).

25. N. J. Finnegan, M. E. Pichard, Magnitude and duration of surface uplift above the Socorro magma body. Geology 37, 231–234 (2009).

26. S. T. Henderson, M. E. Pichard, Decadal volcanic deformation in the Central Andes. Earth Planet. Volcanic Zone revealed by InSAR time series. Geophys. Geochim. Geosyst. 14, 1358–1374 (2013).

27. J. P. Perkins, N. J. Finnegan, S. T. Henderson, M. T. Rittenour, Topographic constraints on magma accumulation below the actively uplifting Uturuncu and Lauzufre volcanic centers in the Central Andes. Geosphere 12, 1078–1096 (2016).

28. J. P. Perkins, K. M. Ward, S. L. de Silva, G. Zand, S. L. Beck, N. J. Finnegan, Surface uplift in the Central Andes driven by growth of the Altiplano Puna Magma Body. Nat. Commun. 7, 13185 (2016).

29. S. L. de Silva, A. E. Mucke, P. M. Gregg, I. Pratome Resurgent Toba—Field, chronologic, and model constraints on time scales and mechanisms of resurgence at large calderas. Front. Earth Sci. 3, 25 (2015).

30. A. Cinque, G. Rolandi, V. Zamparelli, L’estensione dei depositi marini olocenici nei Campi dell’arco in relazione alla vulcano-tettonica. Boll. Soc. Geol. Ital. 104, 327–348 (1985).

31. S. Kauzka, Coastal evolution at a rapidly uplifting volcanic island: Iwo-Jima, Western Pacific Ocean. Quat. Int. 15–16, 7–16 (1992).

32. J. Leuthold, O. Müntener, L. P. Baumgartner, B. Puttlitz, M. Ovtcharova, U. Schaltegger, Time resolved construction of a bimodal laccolith (Torres del Paine, Patagonia). Earth Planet. Sci. Lett. 325–326, 85–92 (2012).
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