Dynamics of episodic magma injection and migration at Yellowstone caldera: revisiting the 2004-2009 episode of caldera uplift with InSAR and GPS data

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Abstract

The 2004-2009 uplift episode is the largest recorded episode of unrest at Yellowstone caldera. We use GPS and InSAR time series spanning 2004-2015, with a focus in the aforementioned event to understand the mechanisms of unrest. InSAR data recorded ~25 and ~20 cm of uplift at the Sour Creek (SCD) and Mallard Lake (MLD) resurgent domes during 2004-2009, and ~8 cm of subsidence at the Norris Geyser Basin (NGB). The SCD/MLD uplift was followed by subsidence across the caldera floor with a maximum at MLD of ~1.5-2.5 cm/yr and no deformation at NGB. The best-fit source models are two horizontal sills at depths of ~8.7 and 10.7 km for the caldera source and NGB respectively, with volume changes of 0.354 and -0.121 km3, and an overpressure of ~0.1 MPa. The InSAR and GPS time series record an exponential increase followed by exponential decrease in the uplift, which is indicative of magma injection into the caldera reservoir, with no need for other mechanisms. However, magma extraction from NGB to the caldera uplift can also be explained by a magma injection model. Distributed sill opening models show that magma is stored across the caldera source with no clear boundary between MLD and SCD. Since the magma overpressure is orders below the tensile strength of the encasing rock, historical episodes of unrest like these are very unlikely to trigger an eruption.

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Key Points:

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7	•	We reanalyze all the ENVISAT InSAR and GPS data that span the 2004-2009
8		episode of unrest
9	•	The GPS and InSAR time series record uplift with an exponential increase
10		followed by an exponential decrease
11	•	Magma injection is the driving mechanism of unrest with no need to invoke
12		exsolved volatiles

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13 Abstract

The 2004-2009 uplift episode is the largest recorded episode of unrest at Yellowstone 14 caldera. We use GPS and InSAR time series spanning 2004-2015, with a focus in the 15 aforementioned event to understand the mechanisms of unrest. InSAR data recorded 16 ~ 25 and ~ 20 cm of uplift at the Sour Creek (SCD) and Mallard Lake (MLD) resurgent 17 domes during 2004-2009, and ~ 8 cm of subsidence at the Norris Geyser Basin (NGB). 18 The SCD/MLD uplift was followed by subsidence across the caldera floor with a max-19 imum at MLD of ~ 1.5 -2.5 cm/yr and no deformation at NGB. The best-fit source 20 models are two horizontal sills at depths of ~ 8.7 and 10.7 km for the caldera source 21 and NGB respectively, with volume changes of 0.354 and -0.121 km³, and an overpres-22 sure of ~ 0.1 MPa. The InSAR and GPS time series record an exponential increase 23 followed by exponential decrease in the uplift, which is indicative of magma injection 24 into the caldera reservoir, with no need for other mechanisms. However, magma ex-25 traction from NGB to the caldera is unable to explain the subsidence coeval with the 26 caldera uplift. The GPS time series of the 2014-2015 episode of caldera uplift can 27 also be explained by a magma injection model. Distributed sill opening models show 28 that magma is stored across the caldera source with no clear boundary between MLD 29 and SCD. Since the magma overpressure is orders below the tensile strength of the 30 encasing rock, historical episodes of unrest like these are very unlikely to trigger an 31 32 eruption.

33 1 Introduction

Silicic volcanoes (SiO₂ > 69%) are responsible for the largest explosive eruptions 34 on Earth (VEI > 8, (*Miller and Wark*, 2008; *Bachmann and Bergantz*, 2008)), more 35 than two orders of magnitude larger than any eruption with recorded visual and in-36 strumental observations. These eruptions form calderas that can remain restless even 37 several hundreds of thousands of years after the climactic eruptions (e.g., (*Hill et al.*, 38 2020)). Several of these calderas undergo transient pulses or cycles of ground uplift fol-39 lowed by periods of either quiescence or ground subsidence ((*Pelton and Smith*, 1979; 40 Dvorak and Berrino, 1991)). However, their relation to potential eruptive activity has 41 remained elusive (e.g., (Pritchard et al., 2019)). The advent of interferometric syn-42 thetic aperture radar (InSAR) geodesy in the early 1990s provided the first detailed 43 images of the spatial and temporal complexities of these ground deformation cycles 44 ((Wicks et al., 1998; Lundgren et al., 2001)), which have been imaged at Yellowstone 45 ((Wicks et al., 1998, 2006; Chang et al., 2007,0)), Long Valley ((Fialko et al., 2001a; 46 Liu et al., 2011; Montgomery-Brown et al., 2015)), Campi Flegrei ((Lundgren et al., 47 2001; Trasatti et al., 2015; D'Auria et al., 2015)) Santorini ((Parks et al., 2012)), La-48 guna del Maule ((*Feigl et al.*, 2014; *Le Mével et al.*, 2015)) and Cordón Caulle ((*Jay* 49 et al., 2014; Delgado et al., 2016,0)) volcanoes. These uplift events have velocities of 50 \sim 1-10 cm/yr, but can reach fast rates up to 28 - 45 cm/yr ((*Feigl et al.*, 2014; *Del*-51 *qado et al.*, 2016)). The spatial and time scales of the deformation events vary from 52 \sim 15 km in Long Valley to more than 70 km at Yellowstone, and from \sim 6 months for 53 Cordon Caulle ((*Delgado et al.*, 2018)) up to at least half a century for Yellowstone 54 ((*Pelton and Smith*, 1979)). These signals have been interpreted as being produced by 55 either magma injection in shallow reservoirs ((Wicks et al., 2006; Delgado et al., 2018; 56 Miller et al., 2017)), volatile exsolution ((Dzurisin et al., 2012; Hildreth, 2017)), fluid 57 flow in the hydrothermal systems that are located in several of these systems ((Hur-58 witz et al., 2007a)) or a combination of these processes ((Dzurisin et al., 2012; Tizzani 59 et al., 2015). However, inherent ambiguities in the interpretation of the geodetic data 60 and the lack of other constraining independent data sets like microgravity, gas chem-61 istry, seismology and heat flow measurements have precluded to unravel the geological 62 mechanism of ground uplift for most of them. Despite the diversity of monitoring data 63 acquired in the past 40 years, recent studies that try to reconcile the wealth of geologic 64 and geophysical data of Long Valley ((*Hildreth*, 2017; *Hill et al.*, 2020)) and Campi 65

Flegrei ((*Troise et al.*, 2019), (*D'Auria et al.*, 2015)) calderas show no agreement upon the driving mechanism of unrest.

The understanding of these unrest signals require a thorough knowledge of the 68 processes that occur inside these magma reservoirs. For instance, all the models avail-69 able for modeling ground deformation data assume injection of fluid magma with 70 Newtonian viscosity into a pressurized cavity ((Lengline et al., 2008; Le Mével et al., 71 2016). This is in contrast with the current understanding of the plumbing system 72 of silicic volcanoes as crystal mushes, in which reservoirs are not molten but solid 73 74 sponge-like bodies with pores filled with interstitial fluids and melt ((Bachmann and Bergantz, 2008; Bachmann and Huber, 2016; Cashman et al., 2017; Cooper, 2017)). 75 These mushes have a protracted grow history by episodic amalgamation of a stack of 76 sill-shaped reservoirs, in agreement with numerical simulations ((Annen, 2009; Annen 77 et al., 2015), and spend most of their lifetime below their solidus under cold storage 78 conditions ((*Cooper and Kent*, 2014; *Rubin et al.*, 2017)). Crystal mushes are unlikely 79 to produce a volcanic eruption unless they are thermomecanically unlocked and remo-80 bilized by many episodic pulses of magma injection ((*Huber et al.*, 2010,0)). However, 81 thermomechanical remobilization is important only over long time scales of 10^2 - 10^3 82 years, while on short time scales of 10^{0} - 10^{1} years magma injection is the principal trig-83 gering mechanism of rhyolitic eruptions ((Huber et al., 2011,0; Degruyter and Huber, 84 2014; Townsend et al., 2019)). Other views indicate that unrest on time scales of 10^{0} -85 10^1 years at large silicic systems may also be explained by melt amalgamation result-86 ing from the inherent instability of buoyant melt layers ((Sparks et al., 2019)). On the 87 other hand, views that consider non-magmatic processes suggest that caldera unrest 88 results from a combination of magma injection, volatile exsolution and/or crystalliza-89 tion and degassing of large magma batches without new inputs of magma. Caldera 90 uplift is then punctuated by episodic leaks of fluids from below the brittle-ductile tran-91 sition (BDT) to shallow areas that deform in a brittle way ((*Fournier*, 2007)). Further, 92 seismic and geodetic data show that episodes of uplift resulting from likely magma in-93 jections are transient features and can be separated by many years ((*Delgado et al.*, 94 2018; Druitt et al., 2019)) or even decades ((Sigmundsson et al., 2010; Druitt et al., 95 2019)) without any other clear evidence for unrest. Other views suggest that caldera 96 resurgence is the direct consequence of episodic magma injection resulting from the 97 incremental and protracted growth of plumbing systems. The episodic uplift is inter-98 rupted by episodes of deflation but the net result is uplift ((Acccella, 2019)). Therefore, 99 regardless of the mechanism of unrest, it is a significant and key question in vol-100 cano science when do these pulses of uplift imply a potential eruption. as it has direct 101 implications for models of hazard assessment (e.g., (*Pritchard et al.*, 2019)). 102

If these uplift events are in turn produced by magma injection, how many of them 103 and of what magnitude are required to actually trigger an eruption? Unfortunately, 104 the models used to study active intrusions ((Lengline et al., 2008; Le Mével et al., 105 2016) do not have predictive capabilities and cannot predict the maximum stress 106 in the reservoir walls produced by magma injection. This is a key element in eruption 107 forecasting models because dikes that transport magma from the reservoir towards the 108 surface form when the deviatoric hoop stress in the reservoir walls reaches a threshold 109 above the tensile strength of the rock which is known to be within \sim 1-40 MPa ((*Tait* 110 et al., 1989; Albino et al., 2010)). Nonetheless, given our imperfect knowledge of the 111 shallow reservoir location, size and physicochemical state, the exact rupture threshold 112 is unknown. Furthermore, the maximum pressurization that reservoirs sustain before 113 an eruption likely varies throughout the lifetime of a single edifice and between different 114 volcanoes ((Lu et al., 2003; Pinel et al., 2010; Carrier et al., 2015)). 115

In this study we focus on the episode of unrest during 2004-2009 at Yellowstone ((*Chang et al.*, 2007,0)), the fastest ever recorded at that volcano since systematic geodetic measurements started in 1975 ((*Pelton and Smith*, 1979)). Yellowstone

caldera has been studied for more than two decades with InSAR and despite the good 119 quality of the geodetic observations, previous studies have used limited amounts data 120 – usually a few interferograms only (e.g., (Wicks et al., 1998, 2006; Chang et al., 2007; 121 Wicks et al., 2020)). Further, despite more than 4 decades of geodetic observations at 122 Yellowstone, there is still significant uncertainty on the driving mechanisms of ground 123 deformation ((Dzurisin et al., 2012; Hurwitz and Lowenstern, 2014)). For example, a 124 detailed conceptual model does not assess the relative contributions of basalt injections 125 and exsolved volatiles ((*Dzurisin et al.*, 2012)). We test the hypothesis of whether the 126 2004-2009 episode of unrest was caused by magma injection or other mechanisms, and 127 particularly the nature of the fluids involved in the episodes of unrest ((*Hurwitz et al.*, 128 2007a; Dzurisin et al., 2012)). To assess these questions, we use all the continuous 129 GPS and all the ENVISAT InSAR data that recorded the complete 2004-2009 episode 130 of uplift with improved source models of ground deformation and solid-fluid mechanics 131 models of magma injection. These models are function of the magma viscosity, magma 132 compressibility and conduit radius among other parameters abd can predict the time 133 series of ground deformation (e.g.m (Lengline et al., 2008; Le Mével et al., 2016; Del-134 gado et al., 2018)). We compare the deformation data and models with other seismic 135 swarms in December 2008 ((Farrell et al., 2010)) and January 2010 ((Shelly et al., 136 2013)) and discuss mechanisms of transition from caldera uplift to subsidence. We fi-137 138 nally extend our models to the most recent periods of unrest during 2014-2015 ((Wicks et al., 2020)) 139

¹⁴⁰ 2 Geological and ground deformation background of Yellowstone caldera

Yellowstone caldera is a $\sim 85 \times 45$ km³ topographic depression and is the youngest 141 of three collapse calderas in the Yellowstone plateau. The eruptions that formed these 142 calderas occurred 2.1, 1.3 and 0.64 Myrs ago erupting the Huckleberry Ridge, Mesa 143 Falls and Lava Creek Tuffs with erupted volumes larger than 2450, 280 and 1000 km^3 144 respectively ((*Christiansen*, 2001)). The last of these eruptions formed the current 145 Yellowstone caldera, which is now filled with 600-1000 $\rm km^3$ of post caldera rhyolitic 146 lava flows. Post caldera volcanism has been focused on the Sour Creek and Mallard 147 Lake domes (SCD and MLD hereafter) (Figure 1) which have been active for the past 148 0.164 Myrs ((*Christiansen*, 2001)). The caldera is underlain by a large plumbing sys-149 tem with large but spatially variable contents of melt (*Farrell et al.*, 2014; *Huang* 150 et al., 2015; Schmandt et al., 2019)). Yellowstone hosts the largest hydrothermal sys-151 tem in the world with half of the world's geysers ((Hurwitz and Manga, 2017)) and 152 several hundreds of hydrothermal vents ((Fournier, 1989; Lowenstern and Hurwitz, 153 2008; Hurwitz and Lowenstern, 2014)). On a geological time scale, the VEI 8 erup-154 tions and the large hydrothermal activity are fuelled by large batches of basalt injection 155 under the upper to mid-crustal silicic system, evidenced by a very large CO_2 degassing 156 flux which requires that the injecting basaltic magma has a CO_2 concentration of 400-157 500 ppm. Such a large amount of CO_2 cannot be dissolved in silicic melts because 158 it would be completely exhausted in 1000 years. Mass balances indicate that ~ 0.3 159 $\rm km^3/yr$ of basaltic melts are intruded beneath the caldera, a similar amount to that 160 intruded at the Hawaii hot spot ((Lowenstern and Hurwitz, 2008; Lowenstern et al., 161 2015)). These injections are also the ultimate source of caldera unrest ((Wicks et al., 162 2006; Dzurisin et al., 2012)). 163

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2.1 Observations and models of caldera unrest

A summary of geodetic observations of ground deformation between 1923 and 2008 is described in detail in (*Dzurisin et al.*, 2012). Ground deformation was observed for the first time in 1975 when leveling lines were measured after 1923 recording 0.7 m of uplift with a time-averaged rate of ~1.4 cm/yr ((*Pelton and Smith*, 1979)). Systematic time-lapse leveling started in 1983 until 2007, and showed that the caldera floor uplifted

until 1984 when the uplift shifted to subsidence following the largest historical swarm 170 ever measured at Yellowstone with M_C magnitudes up to 4.9 ((*Waite and Smith*)) 171 2002)). Caldera subsidence continued until 1996 when a 1 year long episode of caldera 172 uplift was recorded ((*Wicks et al.*, 1998)). Coevally, the area of Norris Geyser Basin 173 (NGB herefater) uplifted between 1996 and 2000 ((*Wicks et al.*, 2006)). Continuous 174 GPS monitoring started in 1996 with a five-fold increase in the station density in 175 2000 (Figure 1). In July 2004, the whole caldera floor uplifted in the largest episode of 176 historic unrest with a maximum uplift rate of 7 cm/yr and with subsidence at rates of 2 177 cm/yr at NGB ((*Chang et al.*, 2007,0)). The caldera uplift ended in mid 2009, coevally 178 with another seismic swarm in the NW part of the caldera (*Shelly et al.* (2013)). More 179 recent episodes of unrest include uplift at NGB between December 2013 and March 180 30 2014, subsidence at NGB and uplift at the caldera between March 2014 and early 181 2015, and NGB uplift and caldera subsidence since then ((Dzurisin et al., 2019; Wicks 182 et al., 2020)) (Figure 1). The transition from uplift to subsidence usually occurs with 183 large seismic swarms at the distal parts of the caldera ((Waite and Smith, 2002; Shelly 184 *et al.*, 2013)). 185

Previous InSAR studies have focused in ERS-1/2 data to measure caldera floor 186 subsidence during 1992 to 1995, slight caldera floor uplift during 1995-1996 and uplift 187 at NGB during 1996-2000 ((Wicks et al., 1998, 2006; Dzurisin et al., 1999; Dzurisin 188 and Lu, 2007; Dzurisin et al., 2012,0; Vasco et al., 2007; Aly and Cochran, 2011; Tiz-189 zani et al., 2015; Wicks et al., 2020)), caldera uplift with ENVISAT during 2004-2009 190 ((Chang et al., 2007,0; Aly and Cochran, 2011; Dzurisin et al., 2012; Tizzani et al., 191 2015)), and NGB uplift, subsidence and then uplift with TerraSAR-X and Sentinel-1 192 data during December 2013 - March 2014, March 2014 - early 2015 and then 2016 -193 2017 respectively ((*Dzurisin et al.*, 2019; *Wicks et al.*, 2020)). Despite the good qual-194 ity of the InSAR observations, all the previous studies have used small amounts of data 195 that provide individual snapshots of the individual episodes of unrest. The only excep-196 tion is (*Tizzani et al.*, 2015) who calculated an InSAR time series with a descending 197 ERS/ENVISAT track for 1992-2010. 198

¹⁹⁹ **3 Deformation results**

We use GPS data from five stations that record the complete sequence of uplift 200 and subsidence during 2004-2009 (Figure 1), operated by the University of Utah and 201 the EarthScope Plate Boundary Observatory (Figure 1, Figure S1). The data were 202 processed by the Nevada Geodetic Laboratory. We use InSAR data from the C-band 203 ERS-1/2, ENVISAT, L-band ALOS and X-band TerraSAR-X satellites (Table 1) pro-204 cessed and analyzed with a variety of methods depending upon the satellite platform, 205 and data temporal resolution (Figure 2 - Figure 3, Figures S2-S3). Data processing 206 follows standard procedures for time series analysis (e.g., (*Doin et al.*, 2011)) and is 207 described in detail in the supplementary material. From the InSAR time series we 208 calculate cumulative total ground deformation during the episode of uplift as the dif-209 ference in deformation between the last image in 2009 and the first image in 2004 or 210 2005. These data span the complete episode of caldera uplift and are hereafter referred 211 as interferograms. 212

Despite the different amount of SAR images and the variable interferogram qual-213 ity of the data in the different ENVISAT tracks, each of the time series record a total 214 of ~ 25 and ~ 20 cm of line-of-sight (LOS) uplift at SCD and MLD between September 215 2004 and September 2009 (Figure 2 - Figure 3). The InSAR data also record ~ 8 cm 216 of subsidence at the NGB between 2004 and 2008 – one year before the end of the 217 uplift at the resurgent domes (Figure 3). However, the onset of deformation at SCD, 218 MLD and NGB cannot be assessed from the InSAR data because there are only two 219 non-winter images in 2004. The wavelength of the deformation signals at SCD, MLD 220 and NGB is constant during 2004-2009 and does not change during the recorded time 221

span, indicating deformation sources that do not change their depth (not shown). The 222 GPS stations OFW2 located near the MLD and HVWY, LKWY and WLWY located 223 near the SCD record between ~ 10 and 20 cm of uplift during the same time span, in 224 agreement with them being at variable distances from the areas of maximum uplift 225 (Figure 1). The deformation signals are similar in location and wavelength than those 226 analyzed in previous studies ((Chang et al., 2007,0; Aly and Cochran, 2011; Tizzani 227 et al., 2015; Wicks et al., 2020)). A seismic swarm that occurred in December 2008 228 and detected by the LKWY station ((Farrell et al., 2010)) is not observed by the In-229 SAR data because we do not include winter images and because the geodetic signals it 230 produced are below the InSAR uncertainty. The caldera uplift transitioned to subsi-231 dence in early 2010 until late 2013. The GPS data recorded this with a constant rate 232 of ~ 1.5 cm/yr, but only the ENVISAT IM2 descending data recorded it (Figure 2C), 233 with an average subsidence of 1-2 cm. ALOS-1 interferograms display double-bounce 234 signals in wetlands that introduce abrupt phase discontinuities (e.g., (Wdowinski and 235 *Hong*, 2015)) and phase unwrapping errors that cannot be corrected. Visual analysis 236 of this data set shows no deformation during 2010-2011, so the data are not consid-237 ered further in this study. The TSX data record no deformation at NGB, and the 238 maximum caldera subsidence at MLD instead of SCD, with a maximum of -3.5 to -2239 cm/yr depending on the track and on the amount of data used in the stacks. The 240 deformation pattern of the 2011-2013 subsidence is significantly different to that of 241 the 2004-2009 uplift. No clear evidence of localized fault creep triggered by magmatic 242 deformation was observed on any of the InSAR time series. Both the GPS data for 243 stations OFW2, HVWY, LKWY and WLWY and InSAR time series during 2004-2009 244 display a pattern of uplift in which deformation increases exponentially until a thresh-245 old is reached and followed by an exponential decrease (Figure 7). This exponential 246 increase followed by exponential decrease is referred hereafter as double exponential 247 ((*Le Mével et al.*, 2015)). 248

²⁴⁹ 4 Source modeling

To understand the sources responsible for the ground deformation at Yellowstone, 250 we jointly invert the interferograms and the GPS vectors with two sources. These 251 include a tensile dislocation ((Okada, 1985)) representing an opening sill below the 252 caldera floor plus an additional source to model the deflation below the NGB – either 253 a pressurized small sphere (McTigue (1987)) or another tensile dislocation. We do not 254 invert the 2010-2012 deformation data because the TSX data show velocity differences 255 up to $\sim 50\%$ with respect to the GPS data. The wavelength of the deformation signals 256 are of several tens of kilometers, suggesting that the deformation sources are likely to 257 lie below the BDT. However, (*Tizzani et al.*, 2015) has shown that viscoelastic effects 258 representative of viscous rheologies are only relevant for time scales longer than 580 259 years which are well below the time span of one decade considered in this study. 260

Prior to source modeling, linear ramps were estimated in areas with no defor-261 mation and removed from the interferograms. The data were then downsampled with 262 a resolution-based algorithm ((Lohman and Simons, 2005)) with a sill geometry at a 263 depth of 15 km. This source is only used to focus the downsampling in areas with 264 deformation and not to enforce an a priori source model (Figure 4, S6). Downsam-265 pling with a shallower sill does not result in a vastly different number of downsampled 266 patches. We use a diagonal data covariance matrix for the InSAR data because the 267 data have a very weak spatial correlation no bigger than a few downsampled pixels and 268 because the far-field variance in non-deforming areas is ~ 7 mm. Both GPS and In-269 SAR data were weighted by the inverse of their uncertainties. Data were inverted with 270 the neighborhood algorithm ((*Sambridge*, 1999)), a non-linear inversion method which 271 iteratively searches for the best-fit model parameters avoiding local minima. Due to 272 the vastly different amount of GPS and InSAR data points – 10 vs \sim 2500 points, the 273

GPS data should be weighted such that the InSAR data will not dominate the best-fit model. Hence, the GPS data were weighted with factors of 1, 0.2 and 0.1 to augment the relative weight of this data set with respect to InSAR and to test the optimal weighting for a joint inversions (e.g., (*Fialko*, 2004)). Inversions with these weighting factors result in models that do not significantly differ from each other, fitting equally well both GPS and InSAR data. Hence, both data sets are assigned equal weights in the non-linear inversion.

The model of a horizontal sill below the caldera floor and a depressurized sphere 281 282 below NGB does not produce a fit as nearly as good compared to that of two dislocations. Therefore we focus on a model of two Okada sills onlys. After convergence was 283 reached by the NA inversion resulting in models that do not significantly differ from 284 each other, we use the Levenberg-Marquardt (LM) algorithm using the NA inversion 285 model as the initial point of this inversion to find the global best-fit model. Inversions 286 for all 14 non-linear model parameters (X and Y sill centroid, depth, strike, dip, width, 287 length) for the two sub horizontal dislocations provide good data fits but fail to con-288 verge to a stable family of solutions because the sources lie on top of each other and 289 thereby they strongly trade-off. After several iterations we fix the dip and strike of the 290 caldera source to 0 and 54 because they converge rapidly to these values. Inversions 291 for the rest of the 12 model parameters converge for the caldera source but not for the 292 NGB source. We discard models in which the NGB and the caldera sills intersect with 293 each other and since the NGB source dip is close to zero, we fix this model parameter 294 to 0. Since convergence was reached for the caldera source parameters, we fix them 295 and then invert for the NGB source, similar to other studies where the deformation 296 signals of different sources interfere with each other (e.g., (*Bagnardi et al.*, 2013)). The 297 best-fit geometry is made up of two horizontal sills (Figure 4, Table 2) at depths of 8.7 298 and 10.6 km for the caldera and the NGB sources respectively (Figure S5). Because 299 we iteratively fixed the model parameters to ensure inversion convergence, it is neither 300 feasible nor meaningful to calculate model parameter uncertainties. To ensure that the 301 model is robust, we also inverted the data with a different algorithm based on a non-302 linear least square iterative inversion ((*Tarantola and Valette*, 1982)) retrieving very 303 similar results. The vertical components of the five GPS stations were inverted for the 304 caldera sill opening and NGB sill closing for every epoch to retrieve the cumulative 305 volume change of the uniform opening model (Figure 6). 306

The Okada model does not include the pressure change as a model parameter 307 so we follow two approaches to estimate it. First, the area of the caldera source sill 308 $(\sim 58 \times 19 \text{ km}^2)$ can be roughly approximated by that of three penny-shaped cracks 309 with a radii a = 9.7 km for each one, and we use the formula $\Delta V = \frac{8}{3}a^3(1-\nu)\frac{\Delta P}{G}$ 310 ((*Fialko et al.*, 2001b)) to get an order of magnitude of the sill pressure change. This 311 approach is just a very coarse approximation and does not imply that an Okada volume 312 change is directly comparable to that of a pressurized penny-shaped crack. Using a 313 volume change of $\Delta V \sim \frac{0.35}{3} \text{ km}^3$ for each of these sources and a shear modulus of 2.1 314 GPa ((*Heap et al.*, 2020)) we get a source overpressure of 0.13 MPa. Second, we fix the 315 caldera sill centroid and dimensions and invert the InSAR data with the DEFVOLC 316 mixed boundary element model (MBEM) ((*Cayol and Cornet*, 1997)) to calculate the 317 source pressure changes produced by a pressurized ellipsoidal crack and a pressurized 318 quadrangle source with a shear modulus of 2.1 GPa ((*Heap et al.*, 2020)). We only 319 invert the IM2 interferogram because it is the most coherent data, several parameters 320 are fixed and to increase the inversion speed. The boundary element model predicts 321 sources at depths of 17 and 13 km respectively and a pressure change of ~ 0.08 MPa, 322 %60 than the value inferred from the crack approximation. These sources are much 323 deeper than those inferred from the inversion of the Okada models. The ovepressure 324 for both models are several orders of magnitude below the tensile strength of the 325 encasing rocks of 10-40 MPa (e.g., *Albino et al.* (2010); ?). 326

The inversion for dislocations with uniform opening results in non-negligible 327 residuals near MLD and SCD (Figure S5), which potentially result from localized 328 areas of fluid pressurization below the resurgent domes. To improve the data fit, we 329 use a distributed sill-opening model for the caldera sill (e.g., (*Delgado et al.*, 2018; 330 *Henderson et al.*, 2017)), in which the best-fit sill is augmented to 12×8 smaller 5×5 331 $\rm km^2$ sills and with the constrain that the sill opening tapers to zero in its edges (Fig-332 ure 4, Figures S6-S7). The model is regularized with Laplacian smoothing to avoid 333 unrealistic oscillatory opening and the amount of smoothing is chosen by the "L curve" 334 corner ((Aster et al., 2018)). We jointly invert GPS and InSAR data with weighting 335 factors α_W between 1 (equal weight for GPS and InSAR), 0.5 and 0.2 to augment the 336 GPS contribution with respect to InSAR. The model fit to the GPS data improves 337 with $\alpha_W = 0.5$ at the expense of a worst data fit to the InSAR data near NGB. Smaller 338 α_W result in a near complete fit to the GPS data but higher residuals for the InSAR 339 data. Therefore, we invert the data with $\alpha_W = 0.5$ which provides good data fits 340 without increasing significantly the residual for the NGB signal recorded by the in-341 terferograms. The distributed opening model predicts volume changes for the caldera 342 source of 0.354 km^3 during 2004-2009 - a time averaged value of $0.07 \text{ km}^3/\text{yr}$, 0.306343 $\rm km^3$ during 2005-2009 and a volume decrease for the NGB source of -0.121 $\rm km^3$ and 344 -0.0981 km³ for the same time periods respectively. The distributed opening models 345 show no clear boundary between the zones of volume change beneath SCD and MLD 346 (Figure 5). A residual of ~ 5 cm is observed in the E part of Yellowstone Lake particu-347 larly in the ascending interferograms and could be related to the ecember 2008 seismic 348 swarm ((Farrell et al., 2010)). 349

³⁵⁰ 5 Dynamic model of magma injection and transport

Kinematic source models like the aforementioned two sills do not provide in-351 sights on the physical mechanism driving the caldera uplift. In this study we focus 352 solely on the mechanism of magma injection from a deep source to a shallow source 353 because there are simple analytic formulas that can be compared directly with ground 354 deformation time series ((Lengline et al., 2008; Le Mével et al., 2016)). Hereafter we 355 refer to magma as molten rock with a Newtonian viscosity. Although the reviews 356 of (Dzurisin et al., 2012) and (Lowenstern et al., 2015) suggest the role of both hy-357 drothermal and magmatic fluids, including exsolved volatiles from cooling magma, 358 in this study we neglect theses effects. This is a clear oversimplification of the very 359 complex hydrothermal-magmatic system of Yellowstone, but it allows to tests to what 360 extent ground deformation can be explained by one of these end-member models. 361

We start with a magma injection model in which the caldera reservoir is con-362 nected to a magma source in the mantle, whose source pressure function increases 363 linearly until a threshold value when it reaches a constant. Magma ascends due to its 364 overpressure and pressurizes the shallow reservoir, resulting in a double exponential 365 function for both the reservoir overpressure and the ground displacement ((Le Mével 366 et al., 2016). In the case of Yellowstone, this model does not take into account the 367 potential connection between NGB and the caldera source, which is addressed later in 368 the study. The magma injection model is defined by Equation 1- Equation 2. 369

$$P(t) = \begin{cases} \frac{st}{t^*} + (s\tau_p - \Delta\rho gL)(e^{\frac{-t}{\tau_p}} - 1) & 0 < t < t^*\\ st(\frac{\tau_p}{t^*}e^{\frac{-t}{\tau_p}} - \frac{\tau_p}{t^*}e^{\frac{-(t-t^*)}{\tau_p}} + 1) & t > t^* \end{cases}$$
(1)

$$\tau_p = \frac{8\eta L V(\beta_w + \beta_m)}{\pi R^4} \tag{2}$$

Here t^* is the transition time between linear increasing and constant deep pressurization, τ_p is a constant that depends on the properties of the plumbing system, R

is the conduit radius, L is the conduit length, V is the reservoir volume, β_w and β_m 372 are the reservoir and magma compressibility and s is the pressurization rate. Since 373 the source pressure is proportional to the displacement for pressurized cavities em-374 bedded in a linear-elastic half-space ((McTigue, 1987)), the model can be scaled with 375 an arbitrary constant to model the GPS time series. This way, the constant scales 376 displacement to pressure for models that do not include the source overpressure as a 377 model parameter (e.g., (*Henderson et al.*, 2017)). Further, if no changes occur in the 378 plumbing system like a change in the source geometry, transient changes in the time 379 series are direct evidence of transient changes in the reservoir pressure and ultimately 380 in the deep source pressure function. As stated earlier, the InSAR data does not show 381 changes in the wavelength of the deformation signals so the sources are fixed in depth. 382 This way we discard that changes in the time series results from changes in the source 383 geometry. 384

The vertical components of the WLWY, LKWY and OFW2 stations are almost 385 insensitive to the opening of the NGB sill (Figure S8), therefore the time series of 386 vertical displacement of these stations are proportional to the caldera sill opening. If 387 the GPS stations record deformation produced by a single source, then the source 388 pressure function is the same for all the GPS stations. The only difference between 389 the time series is the deformation amplitude which is a function of the source geome-390 try and is a constant for each GPS station. Hence the time series can be normalized 391 to account for this constant (Figure 7). The best-fit magma injection model for the 392 normalized vertical component of the WLWY, LKWY and OFW2 stations predicts 393 a transition time of 0.66 years and an exponential time constant of 4 years, with a 394 final adimensional amplitude of 1.37. This implies that the magma injection model 395 predicts ground uplift for at least 5 additional years should inelastic effects be absent, 396 like fluid extraction outside of the caldera ((*Waite and Smith*, 2002)). Inversions for 397 the best-fit magma injection model to the IS2 and GPS time series predict similar time 398 constants and therefore similar properties of the plumbing system (Figure 7a), but a 399 shorter transition due to the lack of InSAR data between September 2004 and May 400 2005. Therefore, the InSAR data are not considered further for these dynamic models. 401 The magma injection model is also applied to GPS time series during 2014-2015, when 402 fast uplift at NGB transitioned to subsidence following a Mw 4.9 earthquake on March 403 30 2014. The model results in good data fits (Figure 8), but the prediction of a sin-404 gle exponential decreasing trend is nearly identical to that of the double exponential 405 model. In this latter model the pressure in the deep magma source is constant during 406 the whole episode ((Lengline et al., 2008)). The InSAR and GPS data and the magma 407 injection model suggest that caldera uplift at Yellowstone during 2004-2009 and 2014-408 2015 is directly indicative of magma injection. Nevertheless, the magma composition 409 cannot be estimated without inferences on the conduit radius (e.g., (*Pedersen and Sig-*410 mundsson, 2006; Fukushima et al., 2010; Delgado et al., 2018)) and the source volume 411 ((Segall, 2019)), the latter not available from the Okada model. The model indicates 412 that magma injection into the caldera source explains the ground deformation with no 413 need for pressurization due to volatile exsolution at the top of the plumbing system 414 or at the bottom of the shallow hydrothermal system. This is in contradiction with 415 an hybrid model of magma injection and volatile exsolution ((*Dzurisin et al.*, 2012)). 416 The magma injection model also ignores the contemporary deflation at NGB, which 417 we address in the following section. 418

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5.1 Two pressurized reservoirs connected to a mantle magma source

The geodetic data show that uplift at the caldera floor is coeval to subsidence at the NGB during most of the 2004-2009 episode of unrest, and that two sub-horizontal sills are responsible for the deformation signals. In this section we use a simple fluidsolid mechanics model based on mass conservation to unravel the potential connection between these two sources of deformation. As our goal is to provide a simple physical model that allows to understand, not model the first order trends observed in the GPS
 time series, we make several geometrical and mechanical simplifications.

In this model, basaltic magma is injected to the shallow caldera source from both 427 the mantle and a deep crustal sources beneath NGB (Figure 9). Although there is clear 428 evidence for lateral transfer and storage of fluids between the SCD and MLD domes 429 (Figure 5), these effects are of secondary order with respect to a single zone of magma 430 accumulation along the two domes. The model of magma flow for two deformation 431 sources embedded in an homogeneous linear elastic half-space is based on a mass 432 433 balance the couples the reservoirs volume and pressure changes with the Poiseuille flow law. These equations are presented in previous studies (*Lengline et al.* (2008); 434 Segall (2013); Reverso et al. (2014); Le Mével et al. (2016); Walwer et al. (2019)) and 435 we adapt them for the specific case of Yellowstone. Since the pressure is proportional 436 to the displacement in a linear elastic half-space, ground deformation follows the same 437 temporal function than the source pressure function. More complex rheologies like 438 viscoelasticity or other mechanisms of fluid transfer such as a flow in a poroelastic 439 media ((Hurwitz et al., 2007a)) are not considered in this study. Also, the model 440 considers neither the mechanical interaction between the sources (e.g., (*Pascal et al.*, 441 (2014)) nor the lateral offset between the sill centroids which are not symmetric and 442 do not lie on top of each other. We assume that since the sources are very large, 443 these boundary effect have a secondary effect. For simplicity we also neglect the short 444 time lag between the onset of inflation at SC and ML and the deflation at NGB. We 445 also neglect the complexity of Yellowstone's plumbing system inferred from seismic 446 tomography ((*Farrell et al.*, 2014; *Huang et al.*, 2015)), including large areas of partial 447 melt, multiphase components in the magma (crystals, dissolved and exsolved CO_2 448 and H₂O). More complex rheologies such as elastic layering, viscoelasticity and plastic 449 effects are not considered. We do consider the effect of magma compressibility due to 450 variations in the reservoir pressure (e.g., (*Rivalta*, 2010)). 451

The volume change rate in the two reservoirs connected with each other, and one of them fed by a mantle magma source is derived from mass conservation and is given by Equation 3 - Equation 4 (Figure 9, e.g., *Reverso et al.* (2014); *Walwer et al.* (2019))

$$\frac{d\Delta M_s}{dt} = \rho_m \frac{d\Delta V_s}{dt} = \rho_m (Q_{in} + Q) \tag{3}$$

$$\frac{d\Delta M_d}{dt} = \rho_{m_2} \frac{d\Delta V_d}{dt} = -\rho_{m_2} Q_{in} \tag{4}$$

with $\Delta M_s, \Delta M_d$ the mass change in the shallow (s) and deep reservoirs (d), 456 ρ_m, ρ_{m_2} the magma density that is injected in the shallow and deep reservoirs, $\Delta V_s, \Delta V_d$ 457 the volume change in the shallow and deep reservoirs, Q the volume flux from a deep 458 mantle source, and Q_{in} the volume flux from the NGB to the caldera source reser-459 voir. Here the shallow and deep reservoirs represent the caldera source and the NGB 460 sill-like sources. The relation between the volume change $\Delta V_{s,d}$ and the resulting 461 reservoir overpressure $\Delta P_{s,d}$ under the assumption that magma is incompressible and 462 the density is constant is Equation 5 463

$$\Delta V_{s,d} = \Delta P_{s,d} \frac{\pi a_{s,d}^3 \gamma}{G} \tag{5}$$

with γ equal to 1 for a sphere ((*McTigue*, 1987)) and $\frac{8(1-\nu)}{3\pi}$ for a penny-shaped crack ((*Fialko et al.*, 2001b)), *G* the shear modulus and $a_{s,d}^3$ the sphere/crack radius. For the general case of volume change due to a pressure change in a reservoir (Equation 6)

$$\Delta P_{s,d} = \frac{\Delta V_{s,d}}{R_{s,d}(\beta_m + \beta_w)} \tag{6}$$

with R the reservoir volume, β_m the magma compressibility and β_w the reservoir compressibility. This results in the Equation 7

$$\Delta V_{s,d} = \Delta P_{s,d} \left(\frac{\pi a_{s,d}^3 \gamma}{G} + R_{s,d} \beta_m\right) \tag{7}$$

The volume flux in a vertical conduit connecting a mantle magma source $\Delta \bar{P}$ to a shallow source ΔP_s is given by the Poiseuille law in Equation 8

$$Q = \frac{\pi a^4}{8\mu H} (\Delta \rho g H + \Delta \bar{P} - \Delta P_s) \tag{8}$$

with *a* the conduit radius, μ the magma viscosity, *H* the conduit length, $\Delta \rho$ the magma-host rock density contrast, *g* the gravitational acceleration and \bar{P} the mantle magma pressure ((*Jaupart and Tait*, 1990; *Lengline et al.*, 2008)). The expression is nearly identical for the conduit connecting the shallow and the deep source in Equation 9

$$Q_{in} = \frac{\pi a_2^4}{8\mu_2 H_2} (\Delta \rho_2 g H_2 + \Delta P_d - \Delta P_s) \tag{9}$$

with H_2 , a_2 and μ_2 the conduit length, radius and magma viscosity in this conduit and ΔP_d the deeper reservoir. Here the flow from the deep reservoir depends upon the pressure gradient ((*Segall*, 2013)) instead of a constant magma flow ((*Reverso et al.*, 2014)). Combining Equation 3 - Equation 9 results in two equations for the pressure change of the two pressurized reservoirs (Equation 10 - Equation 11).

$$\frac{d\Delta P_s}{dt} = \frac{a_2^4 G}{8\mu_2 H_2 a_s^3 \gamma} (\Delta \rho_2 g H_2 + \Delta P_d - \Delta P_s) + \frac{a^4 G}{8\mu H a_s^3 \gamma} (\Delta \rho g H + \Delta \bar{P} - \Delta P_s)$$
(10)

$$\frac{d\Delta P_d}{dt} = -\frac{a_2^4 G}{8\mu_2 H_2 a_d^3 \gamma} (\Delta \rho_2 g H_2 + \Delta P_d - \Delta P_s) \tag{11}$$

We set

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$$\tau_{2}^{-1} = \beta = \frac{a_{2}^{4}G}{8\mu_{2}H_{2}a_{s}^{3}\gamma}$$

$$\tau_{1}^{-1} = \alpha = \frac{a^{4}G}{8\mu Ha_{s}^{3}\gamma}$$

$$\tau_{3}^{-1} = \epsilon = \frac{a_{2}^{4}G}{8\mu_{2}H_{2}a_{d}^{3}\gamma}$$
(12)

If magma compressibility is taken into account, then the time constants become

$$\beta = \frac{\pi a_2^4 G}{8\mu_2 H_2(\pi a_s^3 \gamma + GR_s \beta_m)}$$

$$\alpha = \frac{\pi a^4 G}{8\mu H(\pi a_s^3 \gamma + GR_s \beta_m)}$$

$$\epsilon = \frac{\pi a_2^4 G}{8\mu_2 H_2(\pi a_d^3 \gamma + GR_d \beta_m)}$$
(13)

Arranging terms results in Equation 14 - Equation 15

$$\frac{d\Delta P_s}{dt} = -P_s(\alpha + \beta) + P_d\beta + \alpha(\Delta \bar{P} + \Delta \rho_g H + \Delta \rho_2 g H_2)$$
(14)

$$\frac{d\Delta P_d}{dt} = P_s \epsilon - P_d \epsilon - \epsilon \Delta \rho_2 g H_2 \tag{15}$$

Equation 14 - Equation 15 form a linear system of non-homogeneous differential equations that can be casted in matrix form (Equation 16 - Equation 17)

$$\begin{bmatrix} \frac{d\Delta P_s}{d\Delta P_d} \\ \frac{d\Delta P_d}{dt} \end{bmatrix} = \begin{bmatrix} -\alpha - \beta & \beta \\ \epsilon & -\epsilon \end{bmatrix} \begin{bmatrix} \Delta P_s \\ \Delta P_d \end{bmatrix} + \begin{bmatrix} \beta \Delta \rho_2 g H_2 + \alpha \Delta \rho g H + \alpha \Delta \bar{P} \\ -\epsilon \Delta \rho_2 g H_2 \end{bmatrix}$$
(16)

$$\frac{d\bar{P}}{dt} = G\bar{P} + H \tag{17}$$

with $\bar{P} = [\Delta P_s, \Delta P_d]^T$ the vector that contains the functions for the shallow and 487 deep reservoir pressure. Instead of a piecewise mantle source pressure function of a 488 linear increase followed by a constant after a time threshold ((Le Mével et al., 2016)), 489 we use an exponential function of the form $\Delta \bar{P} = \bar{P}(1 - e^{-\frac{t}{\tau_m}})$ because it is easier to 490 integrate. This function is derived from the data itself (Figure 7) with $\tau_m = 0.36$ years. 491 The solution to Equation 17 is a function of the form $P(t) = \vec{v_1}e^{\lambda_1 t} + \vec{v_2}e^{\lambda_2 t} + \vec{a}e^{-\frac{t}{\tau_m}} + \vec{b}$ 492 with $v_{1,2}$ and $\lambda_{1,2}$ the eigenvectors and eigenvalues of G and the last two terms are 493 vectors derived from the method of undetermined coefficients for the non-homogeneous terms (last term on the right-hand side of Equation 17). The final solution for initial 495 conditions $P_s(0) = 0, P_d(0) = P_{d_0}$ is Equation 18 496

$$\begin{bmatrix} \Delta P_s \\ \Delta P_d \end{bmatrix} = C_1 e^{\lambda_1 t} \begin{bmatrix} 1 + \frac{\lambda_1}{\epsilon} \\ 1 \end{bmatrix} + C_2 e^{\lambda_2 t} \begin{bmatrix} 1 + \frac{\lambda_2}{\epsilon} \\ 1 \end{bmatrix} + \begin{bmatrix} a_1 \\ a_2 \end{bmatrix} e^{-\frac{t}{\tau_m}} + \begin{bmatrix} b_1 \\ b_2 \end{bmatrix}$$
(18)

with the eigenvalues $\lambda_{1,2}$

$$\lambda_{1,2} = \frac{-(\alpha + \beta + \epsilon) \pm \sqrt{(\alpha^2 + \beta^2 + \epsilon^2 + 2\alpha\beta + 2\beta\epsilon - 2\alpha\epsilon)}}{2}$$
(19)

⁴⁹⁸ and constants

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$$b_{1} = \bar{P} + \Delta \rho g H$$

$$b_{2} = \bar{P} + \Delta \rho g H - \Delta \rho_{2} g H_{2}$$

$$C_{1} = P_{d_{0}} - (C_{2} + a_{2} + b_{2})$$

$$C_{2} = \epsilon \frac{(P_{d_{0}} - a_{2} - b_{2})(1 + \lambda_{1}/\epsilon) + a_{1} + b_{1}}{\lambda_{1} - \lambda_{2}}$$
(20)

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Here P_{d_0} is the overpressure produced by injection of basaltic magma at the NGB reservoir during 1996-2000 ((*Wicks et al.*, 2006)). Since basaltic magma is unlikely to significantly cool to produce a significant density change in 4 years, for simplicity we assume that $\Delta \rho = \Delta \rho_2$, $\mu = \mu_2$.

The selected model parameters are in Table 3. Since there are no geophysical 503 constrains on the fourth power of the conduit radius R^4 and the viscosity of the 504 injecting basalt μ , we parametrize the model in terms of the conduit conductivity 505 $\bar{c} = R^4/\mu$ ((Anderson and Segall, 2013)). Basaltic melts have viscosities of $10 - 10^2$ 506 Pa s ((Giordano and Dingwell, 2003)) and conduit flow models during episodes of 507 unrest in basaltic volcanoes show radii of ~1 m ((*Pedersen and Sigmundsson*, 2006; *Fukushima et al.*, 2010)), resulting in $\bar{c} \sim 0.1 - 0.01 \frac{m^4}{Pas}$. The initial pressure with respect to lithostatic conditions are $P_s^0 = 0$ MPa and $P_d^0 = 0 - 0.5$ MPa, the latter 508 509 510 value arising due to the potential magma injection at NGB during 1996-2000 ((Wicks 511 et al., 2006)). The amplitude of the source pressure just scales the pressurization of the 512 reservoirs, so it is not relevant since we are interested on the temporal evolution of the 513 reservoir pressures. We consider cases with and without a density difference, in which 514 the magma ascends due to its overpressure and due to the combined overpressure and 515 buoyancy effects. 516

The simulations show that magma ascends due to its overpressure only (Fig-517 ure 10a-b), predicting the same double exponential pattern for both the caldera source 518 and the NGB source, albeit with a lower amplitude for the latter. However, the 519 model for $P_d^0 = 0$ MPa is unable to predict a double exponential pressurization for 520 the caldera source and a linear depressurization for the NGB source. This is a rather 521 unrealistic scenario as the model predicts that magma ascends from the mantle to 522 the shallow reservoir, and then the high magma overpressure implies that the NGB 523 reservoir must inflate in response to the overpressure, with magma descending 2 km. 524 The effect of magma compressibility is of second order due to the large dimensions of 525 Yellowstone's plumbing system and does not change significantly the sources pressure 526 changes. However, increasing the magma compressibility significantly increases the 527 volume of intruded magma in the reservoir (e.g., Figure 9 in (*Le Mével et al.*, 2016)). 528

On the other hand the simulations show that when magma ascends due to buoy-529 ancy and overpressure (Figure 10c-d), the models predicts pressurization with an ex-530 ponential increase and decrease at the caldera source and both depressurization and 531 pressurization at NGB with near linear trends (Figure 7c-d). We note that the mag-532 nitude of the pressurization of the buoyancy and overpressure model is one of order 533 magnitude larger than the model with magma overpressure only. The only possibility 534 to significantly depressurize the NGB source due to magma flow to the caldera source 535 is to set $P_d^0 = 0 = 0.5$ MPa (Figure 11) and with $\bar{c} \sim 10^0 \frac{m^4}{Pas}$ or smaller but as this 536 value is increased, the magnitude of the subsidence decreases in response to a bet-537 ter hydraulic connection. This is a plausible scenario since it is likely that previous 538 episodes of uplift are inferred to have occurred at NGB ((Wicks et al., 2020)). The 539 simulation predicts subsidence at NGB and double exponential uplift at the caldera, 540 but eventually all the subsidence from material extracted from NGB is counterbal-541 anced by the fluid influx from the caldera source to NGB. The MBEM model predicts 542 $\Delta P = \sim 0.08$ MPa for the caldera source, but it is very unlikely that the NGB reser-543 voir overpressure reached such value during 1996-2000 because the amplitude of uplift 544 during that time span is much lower than the caldera uplift during 2004-2009. This 545 implies than in this scenario the NGB reservoir should have been pressurized decades 546 before 1996, for which we have no quantitative constrains. 547

An alternative model considers that the deep source of magma injection is located below NGB and not below the caldera. Therefore basalt ascends to NGB and then to the caldera, potentially resulting in depressurization during several years at NGB. In this case the second term in the right hand side of Equation 10 must be included in the right hand side of Equation 11 after switching ΔP_s with ΔP_d and modifying the time constant for the NGB reservoir. We also assume a slight overpressure for the NGB source. Changing the force balance with the same pressure function for the mantle source does not result in depressurization for the NGB source for more than 1.5 years until it will start to inflate in response to the incoming magma from the mantle (not shown).

Regardless of the model, none of these simulations can predict at the same time the trends observed in the InSAR and GPS time series at both the caldera floor and NGB and with a constrained set of assumptions available since geodetic measurements started in 1975.

562 6 Discussion

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6.1 Source models and comparison with previous studies

Two families of source models have been proposed for the 2004-2009 episode of 564 unrest: those that rely on horizontal dislocations ((Chang et al., 2007.0; Wicks et al., 565 2020), this study) and those that use a combination of pressure sources ((Aly and 566 Cochran, 2011; Tizzani et al., 2015)). Although both type of models can fit the data 567 well, we consider that the dislocations are more realistic. First, they require less 568 model parameters. Second, neglecting the mechanical interaction between two sills is 569 less inaccurate than neglecting the interaction between at least three pressure sources 570 ((Aly and Cochran, 2011; Tizzani et al., 2015)). Our results are similar to those of 571 (*Chang et al.*, 2007,0) who also found two rectangular dislocations at depths of \sim 7-10 572 and \sim 7-13 km for the caldera and NGB sources, albeit modeling very small data sets. 573 The caldera source model is located at the top of the low V_P zone below the caldera 574 imaged with three-dimensional P wave tomography. This zone has -3 to -4% of V_P 575 difference with respect to the reference velocity model and is inferred to contain little 576 to none partial melt resolvable by this geophysical method ((*Farrell et al.*, 2014; *Huang*) 577 et al., 2015)). Since the magma injection is a discrete event with respect to the spatially 578 and time averaged resolution of seismic tomography, we see no contradiction between 579 the geodetic sources and lack of a clear V_P anomaly. (Wicks et al., 1998, 2006) have 580 argued for two discrete sources of deformation below the caldera floor and episodically 581 active over different times but both the uniform and distributed (Figure 4, Figure S5) 582 opening models indicate that a single dislocation can explain most of the deformation 583 signal during 2004-2009. The caldera source has no clear boundary between the magma 584 accumulation zones below SCD and MLD, except for localized uplift at the SCD, 585 resulting in an additional 5 cm of uplift with respect to MLD. Given the few cycles of 586 deformation observed with detailed geodetic observations, it is not possible to assess 587 if the discrete storage zones below the MLD and SCD ((Wicks et al., 2006, 1998)) are 588 representative of caldera uplift during longer periods of time or not. The NGB source 589 is significantly shallower and different with respect to the source that uplifted during 590 1996-2000 located at a depth of 16 km ((*Wicks et al.*, 2020)), vs 10.7 during 2004-2009. 591 Another difference with respect to (*Chang et al.*, 2007,0) models is that a significant 592 part of the NGB source is located below the caldera floor, and not adjacent to it. 593

Changes in the source geometry can be assessed comparing the location and 594 wavelength of the deformation signals for the different episodes of uplift for data sets 595 that were acquired with the same or very similar flight direction, radar beam and look 596 angle. These data sets include ERS-1/2 descending interferograms and a stack (Figure 597 S4), the ENVISAT IM2 data (Figure 2) and the TSX descending stack. Here the ERS-598 1/2 and ENVISAT IM2 are from the the same track so they have the same line-of-sight. 599 This analysis shows that the wavelength and location of the deformation signals varies 600 during the periods of caldera subsidence in 1992-1995 ((Wicks et al., 1998; Aly and 601 Cochran, 2011), Figure S4), uplift 1996-1997 (Figure 2F in (Wicks et al., 1998)), sub-602

sidence 2000-2002 (Figures 2b-c in (*Wicks et al.*, 2006)), uplift 2004-2009 (Figure 2), 603 subsidence 2010-2013 (Figure 2), uplift 2014-2015 ((*Wicks et al.*, 2020), Figure 8) and 604 subsidence 2015-2020 (Figure 1 in (*Wicks et al.*, 2020)). Deformation at NGB also 605 shows differences in location and wavelength of the deformation signal during the 606 episodes of uplift in 1996-2000 ((*Wicks et al.*, 2006), Figure S4), subsidence during 607 2004-2008 (Figure 2), uplift during early 2014, subsidence during the rest of 2014 (Fig 608 2B in (Wicks et al., 2020)) and uplift during 2015-2019 (Fig 1 in (Wicks et al., 2020)). 609 This implies that the deformation sources are not stable over and they slightly change 610 from one cycle of either uplift or subsidence to the next one. In contrast, other vol-611 canoes show stable deformation sources over several cycles of deformation, even after 612 eruptions ((Lu et al., 2010; Lu and Dzurisin, 2010; Delgado, 2020)). The lack of sta-613 tionary sources indicates patterns of migrating fluids towards shallower depths ((Wicks 614 et al., 2020) and hampers the use of magma dynamics models that rely on a single 615 stable source in depth and location to explain long cycles of unrest (e.g., (*Giudicepietro* 616 et al., 2017)). The spatial variability also indicates a highly dynamic plumbing system, 617 akin to a crystal mush where unrest occurs episodically and in discrete zones of the 618 mush ((*Cashman et al.*, 2017)). On the other hand, do the deformation data indicate 619 a trans-crustal magmatic system in which unrest occurs at multiple depth levels in the 620 crust? The variability in the source depths suggests that this actually occurs at Yel-621 622 lowstone, even on short time scales of less than one year, like during the NGB uplift in early 2014 (Figure 1). However, the exact pattern of fluid migration, potential magma 623 mixing and mingling and stress interaction (e.g., (*Albino and Sigmundsson*, 2014)) are 624 yet to be unravelled. 625

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6.2 Driving mechanisms of unrest

On a geological time scale, the driving mechanism of unrest at Yellowstone 627 caldera are discrete pulses of basalt injection at the base of the rhyolitic plumbing 628 system below the caldera and NGB. (Lowenstern and Hurwitz, 2008) calculated that 629 $\sim 0.3 \text{ km}^3/\text{yr}$ of basalt intrusion with 1 wt% of dissolved CO₂ are required to account 630 for the measured flux of passive CO_2 degassing at Yellowstone, 4 times larger than 631 the time-averaged rate of $\sim 0.074 \text{ km}^3/\text{yr}$ during 2004-2009. A direct comparison be-632 tween these data sets is not possible due to the episodic nature of magma injection 633 and the lack of continuous time-lapse measurements of CO₂ degassing, discussed in 634 detail later. The mechanisms of unrest are less clear over shorter time scales due to 635 the coupling of the shallow hydrothermal system with the deeper magnatic system 636 and volatile exsolution from the injecting basalt $((Dzurisin \ et \ al., \ 2012))$. Also, in-637 dependent data sets and models suggest contradicting mechanisms. Yellowstone lake 638 shorelines have tilted terraces such that the caldera subsidence slightly exceeds caldera 639 uplift during the Holocene ((*Pierce et al.*, 2002)). To account for the slight subsidence, 640 the volume change of exsolved volatiles extracted from the caldera must exceed the 641 volume of injected magma, and these events must alternate in time. This lead (*Pierce* 642 et al., 2002) to suggest that the buildup and extraction of magmatic volatiles is a more 643 likely explanation for the slightly higher subsidence in the Holocene than magma in-644 jection. Further, (*Fournier*, 1989) showed that a crystallizing magma can release 0.026 645 $\rm km^3/yr$ of exsolved fluids from that trapped below the self-sealed layer. This is enough 646 to account for the volume changes that produced the caldera uplift during 1923-1975. 647 However, the hydrothermal model cannot explain the transition from uplift to subsi-648 dence because the seismic swarms that have been recorded coevally to the transition 649 require magma injection. The swarms are likely due to the episodic breaching of a 650 self-sealed at the BDT that leads to fluid extraction from the caldera ((Waite and 651 Smith, 2002)). This process is highly enhanced by a deepening of the BDT, produced 652 by an increase in the strain rate due to episodic magma injection ((*Fournier*, 2007)), 653 contradicting the previous mechanisms for unrest due to volatile pressurization and 654 extraction. 655

Thereby, (*Dzurisin et al.*, 2012) favor a conceptual model that reconciles a wide 656 range of geological, geochemical and geophysical observations. This model suggests 657 that episodic batches of basalt are injected at the base of the rhyolitic crystal mush 658 resulting in reservoir pressurization either at NGB or SCD. As the basalt and the 659 mush crystallize, magmatic volatiles are exsolved. These fluids are in a supercritical 660 state that are trapped below a self-sealed layer in the lower parts of the hydrothermal 661 system and the upper section of the magmatic system resulting in reservoir pressur-662 ization and caldera uplift. The self-sealed layer is also the BDT. Magma injection 663 increases the strain rate, which temporarily deepens the BDT. In this scenario, flu-664 ids in the plastic zone at near lithostatic pressures eventually breach the self-sealed 665 layer, leading to seismic swarms in distal parts of the caldera ((Waite and Smith, 666 2002)), fluid migration outside of the caldera and ground subsidence ((*Fournier*, 1989, 667 (2007)). On the other hand, long-term subsidence at the caldera is likely produced by 668 volatile exsolution from the crystallizing rhyolitic mush that also migrates outside of 669 the caldera ((*Dzurisin et al.*, 1990)). However, (*Dzurisin et al.*, 2012) model does not 670 allow to assess the relative contributions of magma injection and volatile exsolution 671 in the reservoir pressurization (e.g., (*Tait et al.*, 1989)) and the fluids sink sources. 672 Therefore, we compare our results with the previous studies. 673

The LKWY, WLWY and OFW2 GPS fit to the magma injection model (Figure 7) 674 is a strong indication that the driving mechanism of uplift for the caldera source is the 675 injection of $\sim 0.35 \text{ km}^3$ of incompressible basalt during 2004-2009, with no need to argue 676 for exsolved volatiles (discussed later). In this model the pressure of the deep magma 677 source increased linearly until it reached a threshold in early 2005, then it remained 678 constant. This results in a time-variable uplift rate that increased exponentially and 679 then decreased exponentially after 2005 until the hydraulic connection with the deep 680 mantle source was shutdown by inelastic processes (discussed later). Magma is injected 681 in the upper part of the mushy plumbing system inferred from seismic tomography 682 ((*Farrell et al.*, 2014; *Huang et al.*, 2015)). This is also valid for the caldera uplift 683 during 2014-2015 – magma injection at the caldera with source pressure functions that 684 vary from one episode to the next one and a connection between the caldera and NGB 685 sources. On the other hand, there are significant differences. First, the NGB reservoir 686 during 2014-2015 is significantly shallower at a depth of 1-4.5 km ((Wicks et al., 2020; 687 Dzurisin et al., 2019)) vs 10.7 km for 2004-2009, leading (Wicks et al., 2020) to suggest 688 a source of hydrothermal origin. Therefore, we discard that the subsidence at NGB 689 during 2014 would result from magma transfer from this source into the much deeper 690 caldera source, located at a depth of ~ 6 km during 2014 ((*Wicks et al.*, 2020)). Hence, 691 the model of two connected reservoirs cannot be applied to this episode of uplift. We 692 speculate that the reversal from uplift to subsidence at NGB in March 2014 resulted 693 from fluid migration into the shallow hydrothermal system following the breaching of the self-sealed layer that separates the BDT. Since the caldera source did not change 695 its behaviour when NGB uplifted in early 2014 (Figure 8), we speculate that the BDT 696 breaching might have changed the stress field in the deeper source (e.g., (Albino and 697 Sigmundsson, 2014)), potentially allowing for magma to be injected from a mantle 698 source. The exact mechanism is beyond the scope of this study. 699

One significant caveat of the magma injection models is that they do not consider 700 at all the complex structure of Yellowstone's underlying plumbing system inferred from 701 local and teleseismic tomographies ((Farrell et al., 2014; Huang et al., 2015)). These 702 studies show that the volcano is underlain by a low V_p anomaly at depths of 5-17 703 km with 5-15% of melt fraction interpreted as a rhyolitic partial melt underlain by 704 basaltic partial melt. Another low velocity zone is located at depths of 20 to 50 km. 705 with a melt fraction of 2%, extending to the Moho and also interpreted as basaltic 706 partial melt. The two low velocity zones are physically separated. The previous 707 magma injection models neither consider how magma bypasses or interacts in some 708 way with these very large areas of partial melt nor how can the melt segregate through 709

the porous crystalline matrix to ascend through the crystal mush that likely exists in 710 the upper crust. Further, the magma injected during 2004-2009 likely has a basaltic 711 chemical composition compared to that of the mushy rhyolitic reservoir, and they 712 might eventually coalesce on time scales of 10^4 - 10^4 years (e.g., (*Biggs and Annen*, 713 2019)). Magma can also stall somewhere in the crust in a level of neutral buoyancy and 714 undergo viscosity changes resulting from phase transitions. This process could occur 715 deep in the crust such that it might not be detectable with the data. However, the 716 data does not show that the deformation sources change during the episode of uplift. 717 It is unclear from a modeling point of view how the ascending basalt interacts with 718 the plumbing system in the framework of a transcrustal model of unrest on multiple 719 levels in the crust ((Cashman et al., 2017; Sparks et al., 2019; Sparks and Cashman, 720 (2017)). These are all points that have to be addressed in future studies that relax the 721 restrictive assumptions made in the models of magma injection (Figure 9). 722

The spatial coincidence of the SCD with the area of maximum uplift has led other 723 studies to suggest that this is the main area of magma injection ((Wicks et al., 2006; 724 Chang et al., 2007)). Despite the uplift started simultaneously for SCD and MLD 725 in July 2004, we see no clear evidence in the OFW2, WLWY and LKWY stations 726 (Figure 1, Figure S1) to state that magma was first intruded at SCD and then it 727 migrated to MLD, or that magma was injected at MLD and then was stored at SCD. 728 Whatever the situation, this suggests a highly connected area of magma storage that 729 responded coevally to the onset of magma injection with no clear boundary as shown 730 by the distributed sill opening models (Figure 5). GPS observations in the middle of 731 the caldera floor might help to address this point during future episodes of unrest. 732

Magma ascent resulting in reservoir pressurization is due to both its overpressure 733 and its buoyancy with respect to the host rock (Equation 8-Equation 9). The model of 734 connected reservoirs provide insights in this aspect (Figure 10). First, the model with 735 both magma buoyancy and magma overpressure is unable to reproduce the double 736 exponential signals observed at the caldera floor and the NGB, indicating that the 737 signal observed in the GPS time series is indicative of magma overpressure only. This 738 is not a unique characteristic of Yellowstone and has observed at other volcanoes 739 (e.g., (Le Mével et al., 2016)). Second, if buoyancy effects are neglected, then the 740 NGB source inflates in response to the magma flux, resulting in a pressure function 741 742 very similar to that of the caldera source. None of the aforementioned two models predict that subsidence at NGB should end with the waning of uplift at the caldera 743 floor. Whatever the case, the model suggests that magma ascends solely due to its 744 overpressure, and that the subsidence at NGB cannot be explained due to magma 745 extraction towards the caldera source. If NGB is connected to the caldera source, it 746 is not by a mechanism of Newtonian conduit flow. 747

Any mechanism that explains the subsidence at NGB must take into account the 748 very similar onset of uplift with respect to the caldera source, implying that the subsi-749 dence is to same extent triggered by magma injection at the caldera. The hydrothermal 750 system is shallower than 5 km ((*Fournier*, 1989)), so the 2004-2009 NGB source is too 751 deep to be considered of hydrothermal origin. (*Chang et al.*, 2007) explained the rela-752 tion between NGB and the caldera by a mechanism in which the caldera sill opening 753 produces positive dilatation up to 3×10^{-5} strain next to the crack tip. This value is one 754 order of magnitude above the smallest measured strain change produced by dynamic 755 earthquake triggering that can induce transient increases in the medium permeability 756 ((Manga et al., 2012)). This mechanism can induce flow of magmatic volatiles from 757 NGB to the caldera and trigger microseismicity between these two sources. Although 758 the mechanism is plausible, there is a caveat. The NGB sill is located below the BDT 759 and the surrounding medium is plastic with little to none permeability that fluids can 760 use to migrate between the sources. Therefore an opening sill should not produce an 761 increase in the medium permeability because there is no primary porosity ((*Fournier*, 762

2007). The permeability can be increased by magma injection which increases the 763 strain rate, deepening the BDT by ~ 1 km, resulting in a change in the rheological 764 properties and brings deep zones that are plastic into a brittle behavior for a short 765 period of time. However, the NGB and caldera sills are too deep to lie in the brittle 766 region even after the transient increase in the strain rate. Further, lowering the BDT 767 usually results in the breaching of the layer that separates the BDT, not in fractures 768 in zones that are deep into the plastic zone. This implies that the permeability mech-769 anism of magma transport is also not feasible and the connection between NGB and 770 the caldera is uncertain. 771

We have shown that the geodetic signals during the episode of caldera uplift 772 can be explained entirely by magma injection, with no need to invoke volatile ex-773 solution. But this model is neither unique nor necessarily the best explanation. It 774 does not imply that during other episodes of ground uplift magma injection or other 775 mechanisms on unrest can also produce the exact same geodetic signal. For example, 776 rhyolitic plumbing systems are crystal mushes (e.g., (*Bachmann and Bergantz*, 2008)) 777 and Yellowstone's plumbing system has limited amounts of melt ((*Farrell et al.*, 2014; 778 *Huang et al.*, 2015)). Here, exsolved fluids can percolate through the porous matrix 779 and ascend to the top of much where they accumulate in sill-like discrete areas. As 780 the volume of fluids increase this can also result in sill pressurization and ground 781 deformation (e.g., (*Sparks et al.*, 2019)). On the other hand, is it possible that the 782 fluid exsolution, permeable flow and fluid accumulation at the liquid-rich mush cap 783 does not result in detectable pressurization during the caldera uplift? Is it possible 784 that any significant fluid exsolution occurs only in response to the depressurization 785 of the self-sealed layer after it is breached? This seems unlikely. For example, fluid 786 exsolution is not enhanced if the minimum principal stress equals the lithostatic load 787 until the latter equals the pore-fluid pressure (section 10.4.2 in (*Fournier*, 2007)), and 788 this could be attained only after a certain amount of magma has been injected. Fur-789 ther, the mechanisms can vary significantly from an episode to the next one (*Fournier*) 790 (2007)), and mechanisms of unrest that last 10^{0} - 10^{1} years might not be representa-791 tive of the overall caldera behaviour during time scales of 10^4 - 10^5 years (e.g., (*Pierce* 792 et al., 2002)). These scenarios were not considered in this study but are geologically 793 plausible, so future studies should address them. 794

795 796

6.3 Comparison with seismicity, microgravity and stream/gas geochemistry

Ground deformation is one of the several indicators of volcano unrest but uncertainties in the mechanisms that result in ground uplift imply that these data should be analyzed and compared jointly with other independent data sets ((*Pritchard et al.*, 2019)). Here we compare the deformation during the episode of uplift with the dense seismic (e.g., (*Waite and Smith*, 2002; *Farrell et al.*, 2014)) and geochemical ((*Lowenstern et al.*, 2017)) data acquired during more than 30 years at Yellowstone.

Statistics of the number of earthquakes per quarter do not show any abnormal 803 trends during 2004-2008 (Figure 1B in (Chang et al., 2007,0), Figure 8 in (Shelly et al., 804 2013), Figure 1 in (*Farrell et al.*, 2014)). The largest clusters of earthquakes in the 805 caldera during the episode of uplift (Figure 1) occurred during 2004-2006 with micro-806 seismicity located at the northern edge of the caldera floor ((Chang et al., 2007), not 807 shown in Figure 1), and during the December 2008 swarm at Yellowstone Lake ((*Farrell*) 808 et al., 2010)). The seismicity with $M_L > 2.5$ at the onset of uplift is scattered across 809 the caldera with no clear clusters and significantly less than the seismicity triggered 810 when the uplift transitions to subsidence ((Shelly et al., 2013)). Focal mechanisms 811 calculated from waveform first arrivals show normal faulting with seismicity clusters 812 towards the N and S parts of the caldera and with only four events at the SCD ((Russo813 et al., 2017)). (Taira et al., 2010) analyzed five M3+ earthquakes during 2007-2009, 814

two of them the first non-double couple focal mechanisms since monitoring started in 815 1975. These earthquakes are triggered by fluid migration due to an increase in dilata-816 tion from the sill towards shallower opening cracks. (Farrell et al., 2009) calculated 817 the b-value from the Gutenberg-Richter law in a de-swarmed earthquake catalog from 818 1973 to 2006, showing high b-values next to MLD, but no abnormal values indicative of 819 fluid injection at the SCD. In general we conclude that the 2004-2009 caldera uplift was 820 not related to abnormal seismicity in response to magma injection compared with the 821 seismic swarms when deformation shifts from uplift to subsidence ((*Waite and Smith*, 822 2002; Shelly et al., 2013)). This is in contrast with other volcampes, like Long Valley 823 caldera where the onset ground uplift is correlated with increases in seismicity (Fig 3a 824 in (*Hill et al.*, 2020)). The lack of abnormal seismicity during 2004 is not unique of 825 Yellowstone as other volcanoes with very fast deformation, either basaltic like Sierra 826 Negra ((*Davidge et al.*, 2017)), or rhyolitic like Cordón Caulle ((*Delgado et al.*, 2018)) 827 are accompanied by limited amounts of seismicity. One potential explanation for the 828 overall lack of abnormal seismicity is that the plastic rocks around the rhyolitic reser-829 voir cannot be fractured except at the end of the cycles of uplift when the rocks behave 830 in a brittle way for short periods of time. Another alternative is that the 2004-2009 831 cycle of uplift did not produce significant seismicity due to the Kaiser effect ((*Heimis*-832 son et al., 2015)). Here fracturing and seismicity are produced only if the medium is 833 stressed above a previous threshold than in a loading cycle already resulted in frac-834 turing. However, this hypothesis can not be proved as either true or false because 835 only two cycles of uplift in 1923-1984 and 1996-1997 before 2004 were recorded with 836 instrumental observations, both with a significantly worst temporal sampling than the 837 2004-2009 cycle. 838

Micro-gravity data were only measured during 2007-2012 ((*Farrell*, 2014)), and 839 then since 2017 ((*Poland and Zeeuwvan Dalfsen*, 2019)). The 2007-2012 data did not 840 show clear gravity changes but as the data did not include high quality elevation 841 measurements for each gravity station, it did not provide insights on any particular 842 geological process ((*Poland and Zeeuwvan Dalfsen*, 2019)). Therefore the gravity data 843 cannot be directly compared with the InSAR and GPS observations during 2004-2009. 844 (Poland and Zeeuwvan Dalfsen, 2019) measured gravity variations four times during 845 2017 and concluded that the gravity uncertainty of ~ 20 mGal on stable benchmarks 846 is low enough to detect mass changes due to magma injection. 847

In terms of gas and fluid geochemistry, despite the many decades of sampling at 848 Yellowstone's hydrothermal fields (e.g., (Lowenstern et al., 2017)), there is a dearth 849 of long-term geochemical time series. Continuous measurements of CO_2 with eddy 850 covariance instruments have been underway only since 2016 ((*Lewicki et al.*, 2017)). 851 High temporal resolution water chemistry measurements at the Upper Geyser Basin 852 during 2007-2008 ((Hurwitz et al., 2012)) and in major rivers during 2001-2004 and 853 2006-2007 ((*Hurwitz et al.*, 2007b,0)) cannot be compared with the episode of caldera 854 uplift due to their low temporal sampling, or being too distant from the areas of 855 unrest. Decadal time-lapse measurements are only available for chloride discharges 856 in streams with a yearly temporal sampling, but these measurements did not show 857 any unambiguous trend that deviates from the base level values during the period of 858 caldera uplift ((*Hurwitz and Lowenstern*, 2014)). Furthermore, a lateral redistribution 859 of the abnormal chloride flux due to basalt injection would take years to decades until 860 it would result in abnormal fluxes in streams and therefore correlations with ground 861 deformation are not expected to be detected ((Hurwitz et al., 2007b)). These authors 862 also concluded that it would be more feasible to detect perturbations in the shallow 863 hydrothermal system due to deep magma injection by tracking changes in the gas and 864 steam flux instead of the river solute fluxes. Finally, correlations of older episodes 865 of unrest with geochemistry of springs in the NW part of the caldera do not provide 866 meaningful insights ((*Evans et al.*, 2006)). 867

We conclude that the 2004-2009 episode of caldera uplift was not correlated with other large-scale signs of unrest except during the December 2008 dike intrusion ((*Farrell et al.*, 2010)) and the transition from uplift to subsidence in January 2010 ((*Shelly et al.*, 2013)).

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6.4 Eruptive potential of Yellowstone

Eruptions occur when the deviatoric stress in a pressurized reservoir reaches the 873 tensile strength of the rock of $\sim 10{\text{-}}40$ MPa producing a mode I crack that propagates 874 to the Earth surface ((Tait et al., 1989; Pinel and Jaupart, 2003; Albino et al., 2010)). 875 We use the source pressure change as a proxy for the tensile strength of the encasing 876 rock at Yellowstone's magma reservoir. The MBEM inversion and penny-shaped crack 877 approximation predict source overpressures of $\sim 0.08-0.13$ MPa respectively (Figure 878 S7). These values are several orders below the tensile strength of the rock of ~ 10 879 MPa. Therefore individual pulses of uplift like those in 2004-2009 are very unlikely 880 to lead to an eruption. Eruptions should occur every ~ 100 cycles of uplift similar 881 to that of 2004-2009. The recurrence interval of large-scale caldera uplift is not well 882 constrained since geodetic observations with yearly temporal resolution started in 1975. 883 If pressurization cycles last on the order of ~ 5 years, the eruption recurrence is at 884 least 500 years, although there is considerable uncertainty because the caldera uplift is 885 highly transient, with variable magnitudes of magma injection and duration as well as 886 the length of the quiescence periods. If we assume that for every episode of uplift there 887 is an episode of quiescence or caldera subsidence of similar duration, then the eruptive 888 frequency increases to 1 Kyrs. The last eruptions at Yellowstone occurred ~ 70 Kyrs 889 ago ((*Christiansen*, 2001)), which is more than one order of magnitude longer than our 890 eruption interval. If the net record of deformation in the Holocene is slight subsidence, 891 it implies that reservoir deflation is slightly larger than reservoir pressurization and hence the pulses of uplift are even less likely to result in an eruption ((*Pierce et al.*, 893 2002)). 894

Despite the small amount of pressurization, much lower than pressure changes 895 in smaller sills elsewhere (e.g., (Le Mével et al., 2016; Delgado et al., 2016)), a seismic 896 swarm interpreted as a small dike intrusion occurred in December 2008 - January 2009 897 ((*Farrell et al.*, 2010)). The material that intruded the dike was either rhyolitic magma 898 or magma-derived aqueous fluids. The dike is offset with respect to the locus of max-899 imum magma injection at the SCD. This dike produced small displacements recorded 900 by the LKWY station (< than 1 cm, (*Farrell et al.*, 2010)), but the distributed opening 901 model show residuals in the eastern part of Yellowstone Lake that could be explained 902 by this small dike (Figure 4). One explanation is that this intrusion was triggered 903 by reaching the tensile strength of the rock after thousands of loading cycles. Another alternative is that successive cycles of uplift and subsidence at the caldera floor 905 might have decreased and permanently fractured the surrounding rock due to damage 906 loading, effectively lowering the wall rock shear modulus and decreasing the rupture 907 threshold ((Carrier et al., 2015)). If this is true, cyclic pressurization that are unable 908 to trigger a dike intrusion under the aforementioned standard rupture models ((*Pinel* 909 and Jaupart, 2003; Tait et al., 1989; Albino et al., 2010)) can result in small dike in-910 trusions that would not be observed otherwise. Another alternative is these rupture 911 criteria are not valid for the geologic conditions of Yellowstone due to the very large 912 plumbing system of this volcano and its weak crust. Another alternative is that it 913 was not a dike intrusion but only a small swarm (Shaul Hurwitz, personal communi-914 cation). Finally, another alternative is that the swarm was produced by leakage of a 915 916 small amount of magmatic fluids above the BDT, but with no resulting subsidence until the next leakage one year later. Whatever the case, the eruptive potential of 917 Yellowstone deserves a more thorough analysis relating the cyclic loading model with 918 a detailed analysis of the seismic data (e.g., (*Carrier et al.*, 2015)). 919

6.5 Transition of uplift to subsidence

One of the most intriguing features of Yellowstone is the cyclic transition of uplift 921 to subsidence with periods of ~ 10 years (Figure 1). This transition has been explained 922 by the breaching of a self-sealed layer at the BDT due to transient pressurization by 923 either magma injection or exsolved fluids which migrate outside the caldera upon the 924 layer breaching ((*Waite and Smith*, 2002; *Dzurisin et al.*, 2012)). The fluid migration 925 occurs at the end of an uplift cycle and is coeval with seismic swarms in the distal 926 parts of the caldera ((Waite and Smith, 2002; Shelly et al., 2013)). Afterwards, the 927 caldera subsides in response to migration of exsolved fluids from a crystallising mush 928 for several years until another cycle resumes $((Dzurisin \ et \ al., 1990))$. Volumetric 929 changes due to a cooling intrusion produce negligible volume changes and are unable 930 to explain the caldera subsidence ((Dzurisin et al., 1990)). In this model of cyclic 931 deformation, the secular trend of subsidence due to volatile exsolution is reversed by 932 highly transient pulses of basalt injection at the caldera, as in 2004-2009. The Madison 933 Plateau swarm in January 2010 ((*Shelly et al.*, 2013), Figure 1) is interpreted as the 934 breaching of self-sealed layer in the NW part of the caldera, and occurred during the 935 transition from uplift to subsidence. In general, non-eruptive subsidence at volcanoes 936 and calderas like Cerro Blanco ((*Pritchard and Simons*, 2004)), Askja ((*de Zeeuw-van* 937 *Dalfsen et al.*, 2013)) and Medicine Lake ((*Poland et al.*, 2006)) show linear rates \sim 2-3 938 cm/yr over time spans of decades. These have been related to cooling intrusions and a 939 combination of other mechanisms like tectonic extension, but in the absence of a clear 940 sink area, the exact mechanism of subsidence is quite uncertain. 941

The InSAR and GPS observations raise more questions than answers on the 942 mechanisms of caldera subsidence. First, what is the sink of the magmatic fluids 943 extracted from the NGB? (Wicks et al., 2020) suggested that fluids extracted from 944 NGB are injected either in the Norris-Mammoth corridor or the Hebgen Lake fault 945 zone which are the zones with the highest amount of seismicity at Yellowstone. None of 946 the post 2010 data show clear deformation signals outside of the caldera that could be 947 sink sources for some fluids extracted from NGB and the caldera, although the small 948 swath of the TSX data also introduces some uncertainty in this regard (Figure 2). It 949 is also possible that the escaping fluids do not leave a clear geodetic signal if there is 950 no sink reservoir to store them. This is in contrast with the post 2014 deformation 951 which shows small-scale deformation that could reflect fluid pathways outside of the 952 caldera (Fig 7 in (*Wicks et al.*, 2020)). Second, if the end of the caldera uplift is 953 due to some inelastic mechanism of fluid migration above the BDT, then it implies 954 that breaching the self-sealed layer changed the force balance on the conduit that 955 feeds the caldera reservoir. Fluid migration outside of the caldera source implies 956 that the pressure gradient driving this flow is much higher than the pressure gradient 957 between the caldera and the mantle source, suggesting a feedback mechanism. We 958 speculate that the self-sealed layer breaching stopped the connection between the deep 959 mantle and the caldera sources as the magma injection model predicts several years of 960 continuing uplift had the breaching not occurred (Figure 7). From a fluid mechanics 961 point of view, this situation is analog to reservoirs that were inflating prior to an 962 eruption and erupt with reservoir deflation and without co-eruptive magma recharge 963 (e.g., equation 7 in (Segall, 2013), equation S.29 in (Delgado et al., 2019)). Finally, 964 the pattern of cyclic uplift and subsidence (Figure 1) indicates that Yellowstone might 965 behave as a harmonic oscillator with periods of ~ 10 years (e.g., (*Walwer et al.*, 2019)). 966 More geodetic data recorded on future episodes of unrest will shed light in this aspect. 967

968 7 Conclusions

⁹⁶⁹ In this study we have revisited InSAR and GPS time series that span the 2004-⁹⁷⁰ 2009 episode of ground uplift at Yellowstone caldera, the largest instrumentally de-⁹⁷¹ tected at this volcano. Simple solid and fluid mechanics models provide for the first

time quantitative evidence from geophysical data that the caldera uplift results from 972 magma injection from a deep mantle source into a shallow source at ~ 8.7 km. Magma 973 ascent and injection only results from its overpressure, not from buoyancy effects. On 974 the other hand, magma extraction from the NGB source towards the caldera source 975 cannot explain the subsidence that is recorded at the former area. In general, the 976 episode of uplift was only related to small increases in the microseismicity in areas 977 neighboring the caldera, with no clear correlation with other instrumental observa-978 tions. A more complete view of the episodes of unrest can result from a more inte-979 grated view on the different geochemical, geodetic and seismological data sets. Future 980 studies should consider more complex mechanisms of stress-driven fluid migration, as 981 well as a better simulation of the abrupt changes in the force balance that drive the 982 fluid injection and extraction into the NGB and caldera sources. 983

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⁹⁹³ Data Availability statement

ERS-1/2 and ENVISAT data are property of the European Space Agency (ESA) 994 and were provided through GEO Geohazard Supersites and Natural Laboratories and 995 UNAVCO. ALOS-1 date are property of the Japanese Ministerium of Trade and 996 Commerce and were provided by NASA through the Alaska Satellite Facility. The 997 TerraSAR-X data are property of Deutsche Zentrum fr Luft- und Raumfahrt (DLR) 998 and are available at UNAVCO SSARA upon request to Principal Investigator Charles 999 Wicks (USGS). GPS data were provided by Nevada Geodetic Laboratory and also 1000 available by the USGS Earthquake Hazard Program. Earthquakes in Figure 1 are 1001 from the USGS Earthquake Catalog. 1002

1003	Tab	\mathbf{les}

Satellite	λ (cm)	Dates (yyyy/mm/dd)	Pass	\mathbf{Path}	θ	Mode, Beam	#SAR	#Ifg	Atmcor	DEMcor
ENVISAT	5.56	2004/09/03 - 2010/09/17	А	48	19	IM1	26	14	no	no
ENVISAT	5.56	2004/09/22 - 2009/10/21	Α	320	23	IM2	26	42	ERAW2	yes
ENVISAT	5.56	2005/05/24 - 2010/08/31	D	313	19	IM1	12	10	ERAW2	no
ENVISAT	5.56	2005/05/05 - 2010/10/21	D	41	23	IM2	28	37	ERAW2	yes
ALOS-1	23.8	2006/12/30 - 2011/02/25	Α	197	38	FBD-FBS	15	7	linear	no
ALOS-1	23.8	2007/01/16 - 2011/03/14	Α	198	39	FBD-FBS	19	N/A	N/A	N/A
TerraSAR-X	3.1	2011/07/23 - 2013/07/07	Α	45	21	strip_003	12	2*	no	no
TerraSAR-X	3.1	2011/07/28 - 2013/10/19	Α	121	35	strip_009	16	3*	no	no
TerraSAR-X	3.1	2011/07/19 - 2013/10/10	D	159	28	strip_006	10	6*	no	no

Table 1: Details of the processed SAR data sets. The columns show the satellite name, radar wavelength (λ) , date range (year/month/day), whether the satellite is in an ascending (A) or descending (D) orbit, satellite path, average incidence angle (θ) , radar beam except for ALOS-1 where it indicates the radar mode (either FBD or FBS, Fine Beam Double and Fine Beam Single polarization), number of synthetic aperture radar images (SAR) per track, and the number of interferograms used in the time series (Ifg). The asterisk indicates the number of stacked interferograms instead of the number pairs used in the time series inversion. Atmcor is the type of atmospheric correction applied to the data: ERAW2 atmospheric correction with the ERA5 model and an empirical correction with an elevation-dependent term on top of that. DEMcor refers to whether a DEM error correction ((*Ducret et al.*, 2014)) was used or not.

Source model	\mathbf{X}_{s} (km)	\mathbf{Y}_{s} (km)	$\mathbf{Z}_{s}~(\mathbf{km})$	L (km)	W (km)	θ	δ
Sill caldera floor	537.0**	4933.2**	8.7^{**}	57.6^{**}	18.6^{**}	54^{*}	0^{*} 0*
Sill NGB	528.1	4940.0	10.6	22.6	26.6	357	

Table 2: Best-fit sill models. X_s centroid EW coordinate, Y_s centroid NS coordinate, Z_s centroid depth, a major semi axis, b semi-minor axis. Centroid coordinates are in WGS84 UTM 12N datum. Model parameters where iteratively inverted for. First we fixed parameters with * since they converge much faster than any of the others in the NA scatterplots. After many iterations with model parameter convergence we fixed the parameters with **. Finally we inverted the NGB model parameters.

G (GPa)	ν	H(km)	$\rm H_2~(km)$	a_s (km)	$a_d (km)$	γ_s, γ_d	$\Delta ho ~({\rm kg/m^3})$	$V_{R_s} \ (\mathrm{km}^3)$	$V_{R_d} \ (\mathrm{km}^3)$	$\beta_m \ (\mathrm{Pa}^{-1})$
2.1	0.25	10	2	15	10	$\frac{8(1-\nu)}{3\pi}$	270	1000	100	$0-2.1 \times 10^{-10}$

Table 3: Parameters for the analytic model of magma injection connecting two reservoirs. G from (*Heap et al.*, 2020), H_2 and H from the best-fit sill inversion (Table 2), γ_s , γ_s for the crack-like sill reservoirs.

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Figure 1: Top. Yellowstone caldera (thick dashed line), Mallard Lake (MLD) and Sour Creek (SCD) resurgent domes (black lines) draped over the 10m shaded NED DEM. Triangles are GPS stations that recorded data during the complete episode of unrest in 2004-2009. The dots show earthquakes from the USGS Earthquake Catalog shallower than 15 km with $M_L > 2.5$. The red and blue dashed rectangles show the December 2008 ((*Farrell et al.*, 2010)) and January 2010 Madison Plateau ((*Shelly et al.*, 2013)) seismic swarms. Inset shows location of Yellowstone caldera (red triangle) within the United States. The star is the M_W 4.8 earthquake of March 30 2014 at NGB. Bottom. GPS time series of vertical deformation (location on top). The dashed black and grey lines show the 2004-2009 episode of unrest, and the caldera subsidence covered by the TSX data during 2011-2012 respectively. The vertical red and blue lines show the December 2008 and January 2010 seismic swarms.



Figure 2: Mean ground velocities at Yellowstone caldera calculated for the 2004-2009 time period from InSAR time series for ENVISAT data (A, B, C, E, F), and from TSX stacks (D, G). The dashed black line is the caldera border and the thin black lines show the MLD) and (SCD). NGB is the Norris Geyser basin. The black diamonds are the continuous GPS stations used in the study (Figure 1). The color circles, squares and diamonds in (A, B, C, E, F) show the location of maximum uplift for MLD and SCD, and subsidence for NGB in the time series in Figure 2.



Figure 3: ENVISAT time series for selected pixels of maximum displacement at NGB, MLD SCD (Figure 2). The IM2 data show larger dispersion in the displacement because the interferograms used in the time series contain far more turbulent signals than any of the other three tracks (Supplementary Material). The dashed box show jumps in the IM2 descending time series not observed in other data sets. Using pairwise logic, these are not indicative of any ground deformation signal.



Figure 4: Downsampled (A, D), synthetic (B, E) and residual interferograms (C,F) produced by an opening sill below the caldera and by a closing sill below NG spanning 2004-2009. The black and pink arrows are the GPS data and synthetic data from the best-fit joint inversion. The rectangles are the modeled sills (Figure 5).



Figure 5: Distributed sill opening models for 2004-2009 (A) and 2005-2009 (B). The thin and dashed rectangles are the NGB and the caldera sources with uniform opening. The thick dashed line is the caldera border and the elliptical polygons are the SCD and MLD.



Figure 6: Time series of volume change (ΔV) for the caldera floor source with uniform opening, and inverted from the vertical component (U_z) of all the GPS stations. The red line is the best-fit Equation 1, but with P scaled to represent the source volume change.



Figure 7: A. Magma injection model fits to GPS (grey dots) and InSAR (red circles) data. Displacement refers to either vertical displacement for GPS data or line-of-sight displacement for InSAR data. The red and black lines are the best-fit models to the In-SAR IM2 for SCD and the vertical component of the WLWY station. The blue dashed line is the best-fit function $U = U_f(1 - e^{-t/\tau})$ with U_f the maximum displacement and τ a time constant. The vertical lines show the transition between a deep magma source with increasing pressure to a constant pressure and delimit an increasing exponential to a decreasing exponential (green line for GPS and orange line for InSAR). B. Best-fit magma injection model (red line) to the normalized vertical displacements of the WLWY, LKWY and OFW2 stations. The dashed green line is the adimensional mantle pressure function (P_s), and the green continuous line is the best-fit exponential fit of the form $P_s = P(1 - e^{-t/\tau_m})$. The latter function is used to simulate the magma flow between the caldera and NGB (Figure 10 - Figure 11).



Figure 8: Time series of vertical displacement at stations WLWY and NRWY during 2014-2015, with best-fit models of magma injection for a double exponential (solid black line) and exponential fits of the form $U = U_f(1 - e^{-\frac{t}{\tau}})$ (dashed red line). The green dots show the NGB uplift during December 2013 - March 2014. The blue line is a scaled version of the exponential fit but applied to the NRWY vertical component during the same time span. The dashed vertical line shows the transition from uplift to subsidence at NGB coincident with a Mw 4.9 earthquake on March 30 2014. The model fit to the time series indicates two things. First, that the GPS data is indicative of magma injection at the caldera sill. Second, the NGB subsidence was coeval and with nearly the same time history than for the caldera sources. This conincidence was not observed during 2004-2009 (Figure 7).



Figure 9: Sketch that shows the main physical parameters involved in the dynamic model of magma injection (Figure 10, Equation 14-Equation 15). Here P_s and P_d represent the pressure in the caldera floor and NGB reservoirs which are hydraulically connected. Magma ascends from a mantle source to the caldera source, which is also filled by magma flowing from the NGB source. The model does not consider large areas of partial melt inferred from V_P tomography ((*Farrell et al.*, 2014; *Huang et al.*, 2015)) and how the melt can either bypass or interact with these areas.



Figure 10: Simulation of the pressure change in the caldera source (P_s, a, c, e, g) and NGB (P_d, b, d, f, h) reservoirs based on the analytic model of magma transfer between the caldera floor and NGB (Equation 18) for the 2004-2009 time period. Panels a-b) show simulations with conduit flow due to magma overpressure while panels c-d) show models with flow due to both magma overpressure and buoyancy. The color lines show simulations for different conductivities between the caldera sources and a deep mantle source. a-d) and e-h) show simulations for conduct conductivities of 1 and 0.1 between the caldera sill and the deeper NGB sill. Deformation due to magma injection is proportional to the source pressure change in a linear elastic half-space, so the ground deformation follows functions with the same shape than the source pressure function. The models show that magma extraction from NGB to the caldera floor cannot explain both the deformation trends observed in the GPS (Figure 1) and InSAR (Figure 3) data.



Figure 11: Same as Figure 10 but with $P_{d0} = 0.5$ MPa for the NGB sill. This simulation shows that magma extraction from NGB to the caldera can produce subsidence at NGB with a similar amplitude to that of the caldera uplift only if the NGB sill overpressure is unrealistically higher than the overpressure at the caldera sill.

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