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Cubic zirconia in >2370 °C impact melt records Earth's hottest crust

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Bolide impacts influence primordial evolution of planetary bodies because they can cause instantaneous melting and vaporization of both crust and impactors. Temperatures reached by impact-generated silicate melts are unknown because meteorite impacts are ephemeral, and established mineral and rock thermometers have limited temperature ranges. Consequently, impact melt temperatures in global bombardment models of the early Earth and Moon are poorly constrained, and may not accurately predict the survival, stabilization, geochemical evolution and cooling of early crustal materials. Here we show geological evidence for the transformation of zircon to cubic zirconia plus silica in impact melt from the 28 km diameter Mistastin Lake crater, Canada, which requires super-heating in excess of 2370 °C. This new temperature determination is the highest recorded from any crustal rock. Our phase heritage approach extends the thermometry range for impact melts by several hundred degrees, more closely bridging the gap between nature and theory. Profusion of >2370 °C superheated impact melt during high intensity bombardment of Hadean Earth likely facilitated consumption of early-formed crustal rocks and minerals, widespread volatilization of various species, including hydrates, and formation of dry, rigid, refractory crust.

1. Introduction

Shock wave propagation during hypervelocity impact can melt and vaporize both the impactor and target rocks (Melosh, 1989). The immediate, post-impact, thermal pulse far exceeds both ultra-high-temperature metamorphism in tectonic settings (Heisinger and Head, 2006; Korhonen et al., 2014) and the liquidus temperatures of target rocks, and is therefore important for the evolution of planet and asteroid surfaces. The heat associated with impacts had a profound significance on the early Earth, and has been shown to have affected processes in Earth's core (Arkani-Hamed and Ghods, 2011; Monteux et al., 2015; Sleep, 2016), mantle (Watters et al., 2009), crust and atmosphere (Marchi et al., 2016, 2014, 2013; Melosh, 2008).

Volatile depletion and isotope fractionation of Earth, Moon and other planetary bodies are attributed to giant basin-forming impacts (Albarede et al., 2013; Day and Moynier, 2014; Moynier et al., 2010; Pringle et al., 2014). Vaporization of target and impactor material during impacts is temperature dependant. Preferential mass-dependant volatile loss and isotope fractionation from impact melts can therefore locally influence subsequent remelting and rheological behavior, and so giant impacts and/or the cumulative effects of intense periods of impact bombardment have important geodynamic consequences (Albarede, 2009). These effects, in turn, have implications for the evolution and habitability of Earth's surface environment (Abramov and Mojzsis, 2009; Ryder, 2002), and, by extension, the habitability of extraterrestrial planetary bodies and exoplanets.

Initial thermal properties of impact melt also control their ability to digest pre-existing solid material and entrained debris (Onorato et al., 1978), including metals and sulphides essential

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for forming impact-generated economic ore deposits (Keays and Lightfoot, 2004), their cooling rate (Onorato et al., 1978), and consequently the preservation of all pre-existing material and high-P phases formed during the passage of the shock wave (Tschauner et al., 2014). The highest temperature superheated melts (i.e., above the liquidus—the equilibrium temperature for complete melting) and vapor are produced during shock-release melting and frictional melting, whereas exhumation-related decompression melting is limited to sub-liquidus temperatures (Ahrens and O'Keefe, 1971; French and Koeberl, 2010; Melosh, 1989, 2005; Riller et al., 2010; Spray, 1998).

The critical melting temperature of target rocks depends on their composition, proximity to the impact site, initial conditions (Gibbons et al., 1975; Hörz et al., 2005; Stöffler, 1971) and porosity (Kieffer et al., 1976; Wünnemann et al., 2008). Vaporization temperature is taken as the highest vaporization temperature of oxides present in the system; in real systems, incongruent vaporization may occur over a range of temperatures (Ahrens and O'Keefe, 1971; Grieve et al., 1977; Lamoreaux et al., 1987; Lou et al., 1985). However, the complex behavior of silicates has made shock-related heating and, thus, the magnitudes of impact melt superheating and vaporization difficult to predict with any degree of certainty (Ahrens and O'Keefe, 1971). Accordingly, our understanding of temperatures achieved by impact melts necessarily relies on empirical constraints from the rock record via established geothermometers, which have inherently limited temperature ranges (Asimow and Ghiorso, 1998; Cherniak et al., 2007; Hart and Davis, 1978; Kubaschewski, 1982; Lindsley and Andersen, 1983; Sack and Ghiorso, 1991; Taylor et al., 2015). As there are large ranges between liquidus and vaporization temperatures for most rock types, maximum impact melt temperatures are largely unaccounted for, thus the effects of superheated melt may be underappreciated in planetary surface studies.

In the rock record, glass spherules and some types of microtektites contained within impact ejecta horizons provide *a priori* evidence for atmospheric condensation of silicate vapor (Glass and Simonson, 2012). In rare cases, nickel-rich magnesiowüstite inclusions in spherules indicate condensation temperatures >2300 °C (Kyte and Bohor, 1995). However, no direct geological evidence has been reported for high temperatures in contiguous impact melt rock, quenched ejected melt droplets (tektites), or impact melt-bearing breccia (suevite) because no geothermometers applicable to impact melt are calibrated beyond ∼2000 °C (Asimow and Ghiorso, 1998; Cherniak et al., 2007; Hart and Davis, 1978; Kubaschewski, 1982; Lindsley and Andersen, 1983; Sack and Ghiorso, 1991; Taylor et al., 2015). Furthermore, models of impact melt formation often assume initial melt temperatures of ∼1700 °C (Abramov et al., 2013; Onorato et al., 1978; Simonds, 1975), and in some cases allude to temperatures >2000 °C (Wünnemann et al., 2008).

In this study, we investigate the microstructure of zirconia (ZrO_2) produced during the dissociation of zircon ($ZrSiO_4$) in impact glass (quenched impact melt) sampled from the 28-km-diameter, 37.83 ± 0.05 Ma Mistastin Lake impact structure in northern Labrador, Canada (55°53′N; 63°18′W) (Marion and Sylvester, 2010; Sylvester et al., 2013). We present new thermometry based on reconstruction of the polymorphic transformation history of zirconia in a reaction rim around a zircon grain, an approach we refer to as phase heritage. Crystallographic relationships reveal the former presence of cubic zirconia, which definitively quantifies the minimum impact melt temperature to >2370 °C, far exceeding commonly assumed impact melt temperatures. These results thus present fundamental new constraints on the evolution of impact melts, and influence our understanding of the evolution of early planetary crust, when impact rates were orders of magnitude higher than now, substaintial resurfacing by impact structures comparable to or larger in size than Mistastin Lake occurred (Marchi et al., 2014; Morbidelli et al., 2012), and all physical evidence of cratering was subsequently destroyed.

2. Methods and approach

We characterized the zircon grain and surrounding area *in situ* on a polished thin section with cathodoluminescence (CL), backscattered electron (BSE) imaging and wavelength dispersive spectroscopy (WDS). We used electron backscatter diffraction (EBSD) to verify the mineral assemblage and characterize and quantify the crystallographic orientation and microstructure of zirconia (ZrO_2) produced by the dissociation of zircon ($ZrSiO_4$). See Supplementary File for further information about data acquisition and processing. Crystallographic relationships among zirconia grains preserve the phase transformation history, or phase heritage, of zirconia polymorphs (Cayron et al., 2010; Kerschhofer et al., 2000). Transformations between mineral phases commonly occur with systematic crystallographic orientation relationships (e.g., Pearce et al., 2013). New phases tend to nucleate at multiple sites in one of several symmetrically equivalent orientations. The new grain orientations in the resulting polycrystalline microstructure are related by systematic misorientations, described by specific angular rotations around specific crystallographic axes that are strictly controlled by the symmetry relationships between the old and new phases. Therefore, the original (parent) crystal orientation can be inferred by combining the misorientations between the new (daughter) phase grains and the transformation relationships, even when the old crystal has been completely transformed. This concept of orientation-based phase heritage has been used to identify the former presence of high-temperature zirconia polymorphs in manufactured ceramics (Cayron et al., 2010; Chevalier et al., 2009) and kimberlites (Kerschhofer et al., 2000), and the high-pressure $ZrSiO_4$ polymorph reidite in impactites (Cavosie et al., 2016; Timms et al., 2017), and we use it here to investigate the thermal evolution of zircon entrained into impact melt.

The orientation relationships (OR) for the transformations from cubic–tetragonal–monoclinic zirconia due to cooling are well known from the literature on refractory ceramics, and measured orientations of low-temperature monoclinic zirconia (baddeleyite) can be used to reconstruct orientations of parent cubic grains, the highest temperature zirconia polymorph (Cayron, 2007; Cayron et al., 2010). This approach relies on $\langle a\rangle_{cubic} \rightarrow \langle a\rangle_{tetragonal}$ or $\langle c\rangle_{tetragonal}$, generating up to three possible unique tetragonal orientations from a single cubic grain. During subsequent tetragonal–monoclinic transformation upon further cooling, the following transformation rules apply:

$\langle a\rangle_{tetragonal}$ or $\langle c\rangle_{tetragonal} \rightarrow \langle b\rangle_{monoclinic}$

plus either

$\langle a\rangle_{tetragonal}$ or $\langle c\rangle_{tetragonal} \rightarrow \langle a\rangle_{monoclinic}$

(type 1 OR after Cayron et al., 2010)

or

$\langle a\rangle_{tetragonal}$ or $\langle c\rangle_{tetragonal} \rightarrow \langle c\rangle_{monoclinic}$

(type 2 OR after Cayron et al., 2010)

allowing for up to four unique orientation variants from each tetragonal identity (Chevalier et al., 2009), totaling twelve possible monoclinic variants from a single cubic identity (Cayron et al., 2010).

The Python-based software combination of ARPGE and GenOVa were used to perform a crystallographic orientation analysis of baddeleyite (monoclinic zirconia) to reconstruct evidence for precursor zirconia polymorphs following the approach applied by Cayron (2007) and Cayron et al. (2010) for ceramic applications. This method automatically and objectively reconstructs 'parent' grains using known sets of orientation relationships among 'daughter' grains that have been established to result from phase transformations between specific zirconia polymorphs.

3. Results

The zircon grain analyzed, MZRN-1, is in a glassy impact melt rock, exposed on top of the 80 m thick columnar jointed impact melt body at Discovery Hill near the crater wall (Marion and Sylvester, 2010). Mistastin impact melt rock composition ranges from ~53 to ~59 wt% SiO_2 and is interpreted to reflect mixing of different proportions of various crystalline igneous target rocks (Marion and Sylvester, 2010). Field relations indicate the melt ponded at Earth's surface at ambient atmospheric pressure (Marion and Sylvester, 2010). The zircon records a pre-impact U–Pb crystallization age of 1403 ± 10 Ma (2σ error, $n = 3$, MSWD = 0.85), consistent with derivation from Mesoproterozoic bedrock in the region (Zanetti, 2015). Other phases in the matrix include ballen-textured silica, which is also characteristic of high-temperature (>1200 °C) impact melt (Ferrière et al., 2009).

Zircon grain MZRN-1 has an oscillatory zoned core in cathodoluminescence (CL) that is truncated by a ~2 μm wide, bright CL rim that contains zirconia particles with interspersed sub-μm silicate melt inclusions (Fig. 1). The grain is cut by irregular fractures and contains two voids, all filled with glassy silicate impact melt (Fig. 1B). The zircon does not preserve any diagnostic shock-deformation microstructures, such as twins or the high-pressure $ZrSiO_4$ polymorph reidite (Timms et al., 2017) (Supplementary File item 2). Minor lattice misorientations (<5°) are associated with rigid block rotation across brittle fractures (Supplementary File item 2), but in general the core of the zircon appears undeformed.

The core is surrounded by a ~40 μm thick corona of vermicular ZrO_2 crystals ranging from ~0.5 to ~5 μm across and up to ~20 μm long (Fig. 1). Elongate crystals tend to be aligned at high angles to the zircon core, and form domains of morphologically similar clusters up to ~50 μm wide. Individual crystals generally do not impinge on each other, and are separated by glassy silicate impact melt similar in composition to the surrounding crystal-free impact melt (Fig. 1). ZrO_2 contains lower trace elements abundances (e.g., ~350 ppm Ti, ~275 ppm Th, ~90 ppm Y) than zircon (e.g., ~900 ppm Ti, ~600 ppm Th, ~1900 ppm Y) (Supplementary File item 4). Most (>99%) of the ZrO_2 grains index as baddeleyite (monoclinic ZrO_2). No crystalline SiO_2 phases were detected (Figs. 1, 2).

Baddeleyite grains are commonly twinned and have a wide range of crystallographic orientations (Fig. 2A, C). Morphologically distinct clusters of baddeleyite grains preserve up to twelve unique crystallographic orientation variants with a systematic relationships among them (Fig. 2D, Supplementary File item 5). Groups of grains are crystallographically orientated approximately orthogonal to one another, and systematic deviations from orthogonality of ~20° form cross-shaped patterns on pole figures of ⟨100⟩ (Fig. 2D).

The occurrence of spatially-clustered baddeleyite grains with these distinctive patterns of twelve orientations uniquely identifies the former presence and orientation of a precursor cubic zirconia polymorph from the two-stage transformation from cubic to tetragonal to monoclinic zirconia (Fig. 2B, E) (Cayron et al., 2010). Initial processing of the EBSD maps of MZRN-1 involved dilation of daughter monoclinic zirconia grains by an iterative nearest neighbor extrapolation routine so that they impinge on one another (Supplementary File Fig. S3). This was required in order to perform neighbor-pair disorientation analysis within the ARPGE software.

Observed peaks we identified in disorientation analysis of MZRN-1 at 90°, 115°, and 180° are consistent with cubic–monoclinic transformation twinning (Fig. 3). We identified symmetry operators for disorientations between daughter monoclinic zirconia grains at these specific angles in crystal reference frame (Fig. 3).

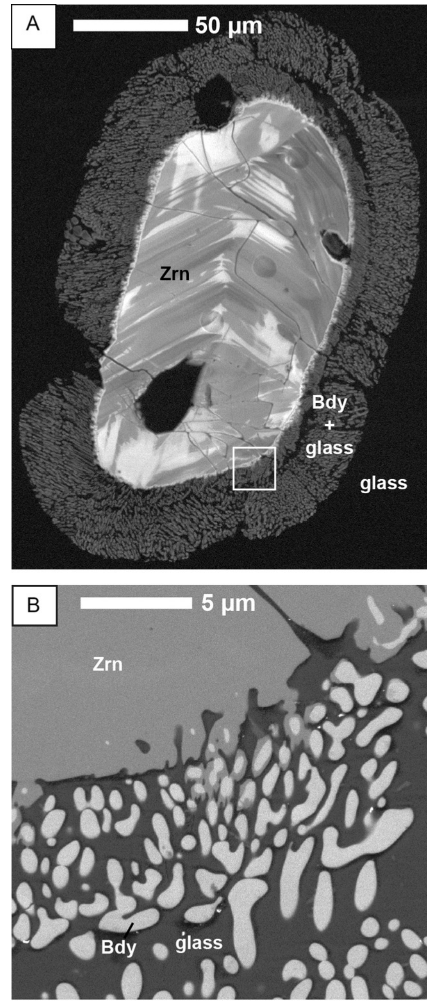

Fig. 1. A. Panchromatic cathodoluminescence (CL) image of zircon (Zrn) xenocryst with a corona of baddeleyite (Bdy) and silicate glass. B. Backscattered electron image of area shown by white box in A showing the zircon–baddeleyite–glass interface.

We used the software GenOVa to develop a list of theoretical symmetry operators for the cubic–monoclinic transformation according to OR type 1 or type 2 OR (Cayron et al., 2010) to reconstruct the parent cubic zirconia grains (Figs. 2, 3). Parent grain reconstruction involved a quadruplet search method with a minimum of five daughter grains per parent grain (Cayron et al., 2010). The disorientations we observed (angles + axes) fit very well with rotations by the operators expected from OR type 2, within an angular tolerance range of <5° (Fig. 3). We conclude that OR type 2 is the OR in the samples: there is no evidence that OR type 1 is present. The observed spread of the orientations (Fig. 3) is not due to a mixture of OR types 1 and 2, and may be due to limited crystal-plasticity in the dynamic impact melt environment. We used 15–20° as the tolerance angle for the parent cubic grain reconstruction (Fig. 2B, D, Supplementary File Fig. S5).

Reconstruction of parent cubic zirconia grains from baddeleyite electron backscatter diffraction (EBSD) data shows that the dissociation corona was once comprised of large (~5 to ~50 μm across) domains of vermicular single crystals of cubic zirconia (Fig. 2B). The morphology of the cubic parent grains indicates that the vermicular ZrO_2 grains formed a connected 3D network that broadly pseudomorphed the original rim of the zircon grain. A second zir-

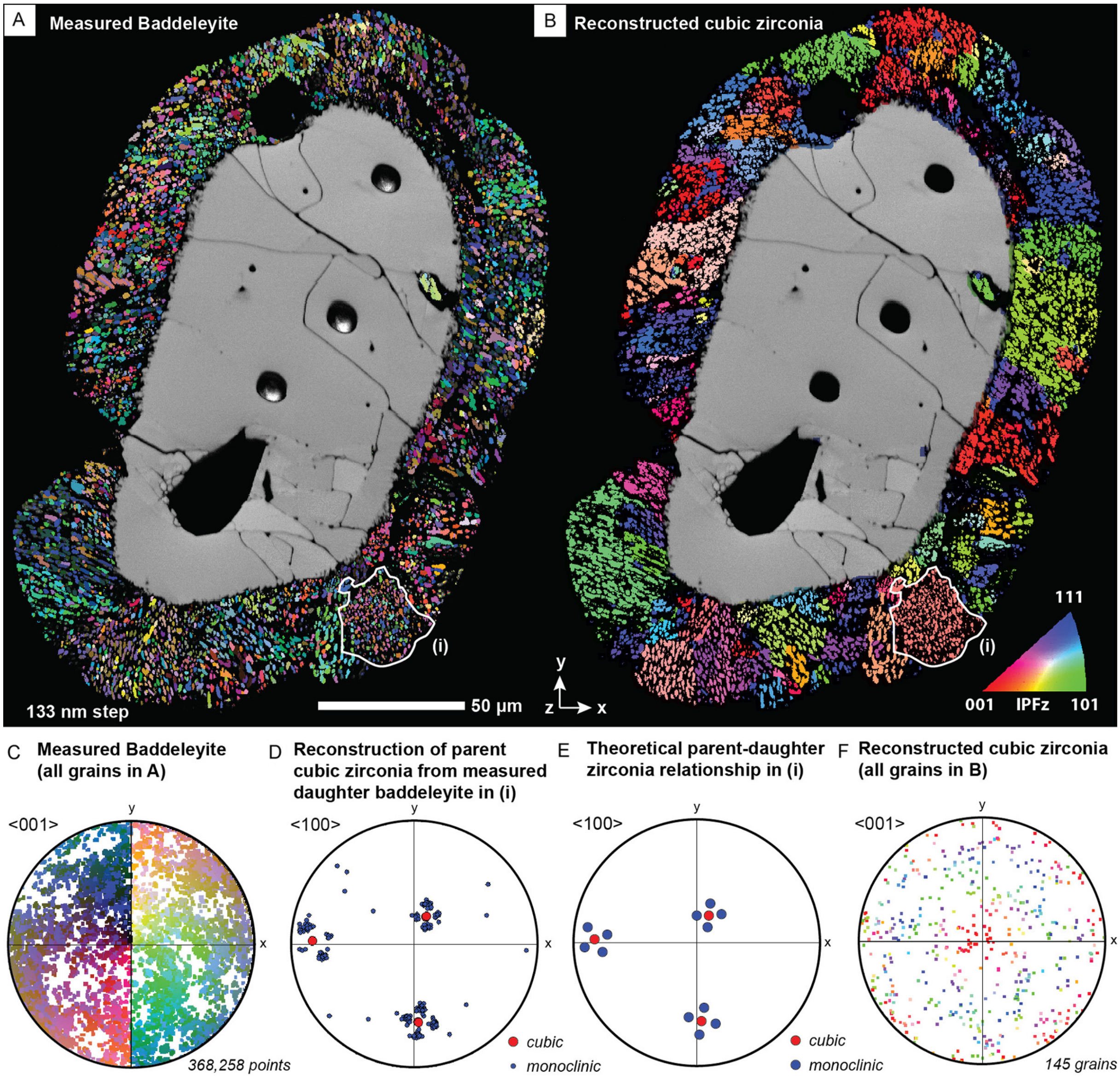

Fig. 2. A. Electron backscatter diffraction map colored for measured crystallographic orientation of baddeleyite. Zircon is shown in grayscale based on EBSD pattern quality (band contrast). B. Map showing crystallographic orientations of reconstructed parent cubic zirconia grains. Inverse pole figure (IPF) orientation color scheme. C–F. Pole figures showing measured 'daughter' baddeleyite orientations and reconstructed cubic zirconia 'parent' orientations. Lower hemisphere plots in sample *x*–*y*–*z* reference frame. (For interpretation of the colors in this figure, the reader is referred to the web version of this article.)

con from the same sample (MZRN-2) displays the same partial dissociation texture and also successfully yields reconstructed parent cubic zirconia grains (Supplementary File items 3 and 5), indicating the same cubic zirconia phase heritage as MZRN-1.

4. Discussion

The microstructural and geochronological data, field relations, and thermodynamic constraints all support the interpretation that the zircon grains described here record an extreme high-temperature history at low-pressure conditions. The Mesoproterozoic U–Pb age and preserved igneous zoning of MZRN-1 indicate that the grain is a xenocryst derived from one of the plutonic igneous target rocks. The complete lack of crystal–plastic deformation, shock-twins and high-pressure polymorphs are consistent with the grains not having been affected by high-pressure shock conditions. The grains were, however, entrained into impact melt in the dynamic crater environment after passage of the shock wave. The ZrO_2 coronae are therefore interpreted to have developed solely due to thermal effects from immersion in super-heated impact melt at ambient pressure. The inference of ambient pressure is also consistent with available thermodynamic constraints that indicate that the zircon dissociation reaction line has a steep Clapeyron slope (29 °C/bar) and therefore is inevitably a low-pressure process (Timms et al., 2017).

A minimum temperature achieved by the impact melt can be constrained via analysis of phases in the zirconia corona, and we use the ZrO_2–SiO_2 binary phase diagram at 1 atm to establish the temperature of zircon dissociation (Fig. 4) (Kaiser et al., 2008). The presence of a thick zirconia corona clearly records the melt achieving high enough temperature for zircon dissociation to proceed (Fig. 4) (Kaiser et al., 2008). Phase equilibria predict that zircon dissociates to tetragonal zirconia and cristobalite at 1673 °C (Fig. 4). The presence of silicate melt and absence of cristobalite indicates the sample exceeded 1687 °C; liquid silica released by dissociation would have dissolved into surrounding melt.

Clusters of morphologically distinct zirconia crystals that contain up to twelve unique orientations with systematic misorien-

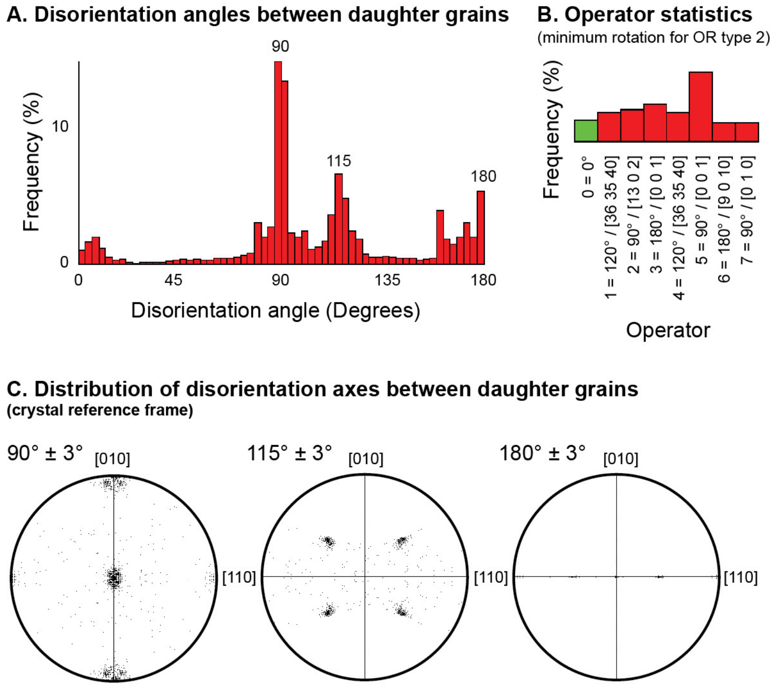

Fig. 3. Results from orientation analysis of daughter monoclinic zirconia grains of MZRN-1 using the software ARPGE (Cayron, 2007; Cayron et al., 2010). A. Histogram of disorientation angles between daughter monoclinic grains. B. Disorientation symmetry operator statistics for adjacent daughter grains for operators for the type 2 orientation relationship (OR) of Cayron et al. (2010). C. Pole figures showing the distribution of 90°, 115°, and 180° disorientation axes between daughter grains in the crystal reference frame.

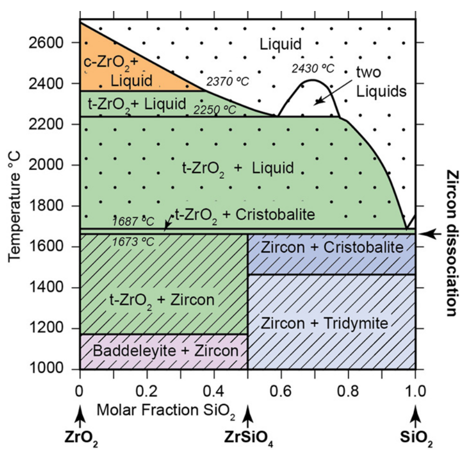

Fig. 4. T–X phase diagram for ZrO_2–$ZrSiO_4$–SiO_2 system modified after Kaiser et al. (2008). Tridymite stability after Swamy et al. (1994). Crosshatch shows zircon-bearing fields, stipple shows melt-bearing fields.

tations are consistent with a two-stage solid-state transformation from cubic zirconia via tetragonal zirconia. The former presence of cubic zirconia requires that the zircon, and therefore the melt that surrounds it, reached a minimum temperature of 2370 °C, but did not exceed 2700 °C above which cubic zirconia would have melted (Fig. 4), and thus prevented crystallographic phase heritage relations from being preserved. Upon cooling below 2370 °C, the displacive transformation of cubic to tetragonal zirconia results in up to three distinct orientations of tetragonal zirconia whereby $(001)_{tetragonal}$ is parallel to a $\{100\}_{cubic}$ (Heuer et al., 1987). Further cooling below ∼1180 °C results in a stable assemblage of baddeleyite + zircon (Fig. 4). The tetragonal–monoclinic zirconia transformation is martensitic, and resulted in four orientation variants of baddeleyite from each tetragonal parent (Cayron et al., 2010). Thus, the transformation of higher-temperature cubic ZrO_2 polymorphs in the corona to the low temperature ZrO_2 polymorph baddeleyite produced systematic orientation relationships among baddeleyite grains. These are interpreted as transformation (reversion) twins formed upon cooling (Supplementary File item 6).

Even though cubic zirconia is the highest temperature ZrO_2 polymorph, its stability is pressure-dependant, such that it is stabilized at lower temperature with increasing pressure (Bouvier et al., 2000; Kerschhofer et al., 2000) (Supplementary File item 7). However, zirconia in the Mistastin sample is a product of zircon dissociation, which ostensibly is a low-pressure phenomenon that occurs at post-shock conditions (see Timms et al., 2017, for further details and thermodynamic calculations). All available field evidence, theoretical constraints, microstructural and geochemical data supports zircon dissociation in an unconfined impact melt at ambient, post-shock pressure. Therefore, the transformations among zirconia polymorphs that took place after zircon dissociation necessarily also occurred in the melt sheet at post-shock low-pressure conditions, which constrains the stability of cubic zirconia to extremely high temperatures. Impurities, such as REE and Y, can also stabilize cubic zirconia at lower temperatures (Swab, 2001) (Supplementary File item 7). However, the levels of trace elements present in the ZrO_2 are much less than 0.1 wt% (Supplementary File item 4), and so the effects of impurities on the cubic to tetragonal zirconia transformation temperature were negligible (Swab, 2001) (Supplementary File item 7). Therefore, the former presence of cubic zirconia in the Mistastin Lake zircon indicates that the impact melt temperature was in excess of 2370 °C (Fig. 3) (Kaiser et al., 2008).

Our study documents forensic geological evidence for a minimum temperature of 2370 °C in a sample of impact melt. This temperature constraint is the highest recorded by any rock on Earth's surface, and is several hundred degrees higher than previous estimates for average temperatures of superheated impact melts (Fig. 5) (Onorato et al., 1978; Simonds, 1975; Wünnemann et al., 2008).

The ability to assess the phase heritage of ZrO_2 polymorphs by crystallographic orientation relationships is a new methodology for establishing the thermal evolution of superheated melt. Tex-

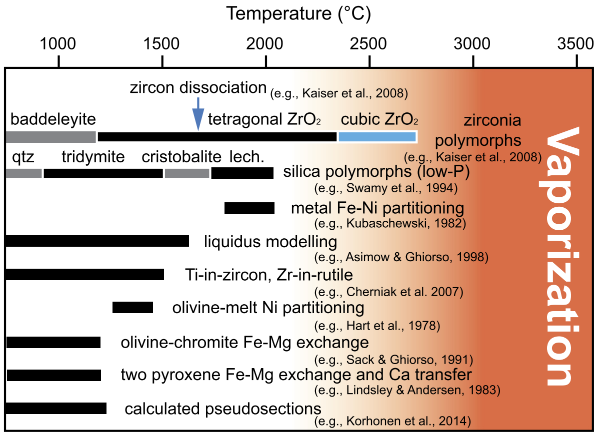

Fig. 5. Ranges for various types of high-temperature thermometers available in geosciences. Cubic zirconia (blue bar) related to zircon dissociation (blue arrow) is the thermometer for highest known temperatures (numbers on left indicate references). Vaporization field is approximate and based on rates for common species present in silicate melts (Lamoreaux et al., 1987), qtz = quartz; lech. = lechatelierite.

tural evidence of zircon dissociation and associated zirconia has been reported in numerous examples, including impact melt rocks, tektites, lunar melt breccias, kimberlites, as well as anthropogenic settings, such as slag from metal smelting and nuclear test sites (Timms et al., 2017). Given that zircon is a common accessory mineral in many rocks, this approach can be used to assess minimum temperatures of zircon-bearing impact melt rocks from Earth, Moon, and meteorites. The broader use of determining phase heritage can potentially be applied to a wide range of polymorphic transformations in geological systems.

Our results begin to constrain the conditions under which zircon and its U–Pb isotope systematics may survive during impact cratering. During high temperature and low shock pressure events, zircon is capable of surviving extreme conditions without necessarily recording microstructural evidence typical of shock deformation (Erickson et al., 2017; Timms et al., 2017, 2012). The Mesoproterozoic U–Pb protolith zircon age preserved in the ~38 Ma impact melt investigated here indicates that, despite being entrained in impact melt in excess of 2370 °C, the residence time at high temperature was not sufficient to cause significant Pb-loss from the core of the grain by volume diffusion (Cherniak and Watson, 2001). The glassy matrix is further evidence that duration at extremely high temperatures was short and the sample quenched rapidly. However, increased melt volumes in larger impact structures with slower cooling rates may lead to complete digestion of zircon xenocrysts, erasing mineralogical evidence of superheated temperatures. Early during Earth's history when impact rates were high, digestion of zircon by superheated impact melt may have contributed to the paucity of Hadean zircon preserved in the geological record (Marchi et al., 2014).

The formation of >2370 °C impact melt during moderate sized impacts has several significant, global-scale consequences during the early evolution of the Earth and Moon when the frequency and magnitude of meteorite flux was much higher (Gomes et al., 2005; Kring and Cohen, 2002). The extent of compositional modification by differential volatile loss of major and trace elements from impact melts by selective vaporization at temperatures >2370 °C has not been fully investigated, yet are sufficiently high to cause significant volatilization of many key oxide constituents of silicate melts (Lamoreaux et al., 1987). Preferential volatilization of SiO_2, Na_2O and K_2O components from hot impact melt bodies could provide a mechanism for localized densification of residual crust, potentially with geodynamic consequences. Molecular species essential for terrestrial life, such as H_2O and CO_2, are potentially lost due to volatilization of superheated impact melt. High-temperature impact-melting could locally form dry, rigid, refractory residual crust, resistant to subsequent melting. Moderate sized impacts, comparable to Mistastin, are predicted to have completely resurfaced the Earth within the first tens of millions of years after the Moon-forming event, irrespective of the choice of the suggested bombardment models (Marchi et al., 2014). Therefore, the resultant net effect of hypothesized high impact flux early during Earth's history, combined with commonplace high temperatures of impact melts, potentially led to widespread depletion of hydrous minerals from the lithosphere and therefore governed rheological properties of early crust. Further investigation of the geodynamic consequences of this phenomenon for early Earth would require accurate predictions of impact melt temperatures, flux, size, impact angles, and velocity distributions of impactors, and appropriate scaling relationships for melt volumes linked to these variables. Furthermore, long term effects would need to account for rehydration effects by interaction with Earth's hydrosphere. Our results on high temperature melt and attendant volatilization processes have implications for the early evolution of the geochemical reservoirs and multi-element isotope systematics of planetary bodies (Drake and Righter, 2002) and that extremely high melt temperatures can be achieved even in moderate-sized impact events, and are not limited to giant, basin-forming impacts.

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Appendix A. Supplementary material

Supplementary material related to this article can be found online at http://dx.doi.org/10.1016/j.epsl.2017.08.012.

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