This Supplemental Material accompanies

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SUPPLEMENTAL MATERIAL

1: Mexican terranes as sediment sources

The continental mass surrounding the GoM comprises several heterogeneous terranes with dissimilar origin. Figure 1 displays the distribution of juxtaposed Laurentia- and Gondwana-related terranes in Mexico. Their geological history includes multiple events of collision, lateral movements, rifting- and subduction-related magmatism, thermal episodes and orogenies since the Precambrian, from the breakup of Rodinia through Gondwana and Pangaea assembly and breakup until the present day.

Based on isotopic signatures of igneous rocks and the surface distribution of basement outcrops, Centeno-García (2017) compiled an updated classification for the tectono-stratigraphic units, or terranes, of Mexico after their paleogeographic affiliation in three general categories: 1) Laurentian terranes, distinguished by Precambrian basement or old crustal signatures in the isotopic composition of their younger igneous rocks; 2) Gondwanan terranes, characterized by Precambrian, high-grade metamorphic basement or lower- to mid-Paleozoic metamorphic basement (mostly schist complexes); and 3) Pacific terranes, characterized by Paleozoic to Mesozoic basement, or igneous and metamorphic rocks that show juvenile isotopic signatures.

The opening of the GoM occurred in the Early-Middle Jurassic (e.g., Pindell, 1994; Pindell and Kennan 2009). Since then, sediments have accumulated in this basin transported by fluvial systems as the product of erosion of exposed rocks. The tectonic evolution and exhumation history of the Mexican terranes are critical controls of the sedimentary fill

accumulated in the GoM. Mexican terranes are composed of heterogeneous basement units that are, ultimately, the source of clastic sediments to the GoM basin. Moreover, we have to emphasize the importance of other clastic sources such as the Mesozoic and Cenozoic magmatic arcs (e.g., Trans Mexican Volcanic Belt TMVB), which developed as the result of subduction of the Pacific and Cocos plates under the Mexican mainland (e.g., Ferrari et al., 2012). These potential source areas are depicted in Figure 2.

The southern part of California and northwestern part of Mexico is formed by a large igneous body emplaced at and near the continental margin during the early history of the so-called Peninsular Ranges Batholith, which occupied the western margin of mainland Mexico prior to the Miocene opening of the Gulf of California (e.g., Dickinson and Lawton, 2001). Lower Cretaceous igneous rocks are widespread in southern California and the northern half of the Baja Peninsula, where they include scattered plutons (134–128 Ma) and volcanic rocks (128–115 Ma) assigned to the Santiago Peak volcanics in the north and the Alisitos Formation in the south (Wetmore et al., 2003). This Early Cretaceous volcano-plutonic complex, broadly termed the Alisitos arc, has been interpreted as a continental-margin arc developed on accreted oceanic material of the Guerrero volcanic terrane (e.g., Dickinson and Lawton, 2001) or, for most of its extent in the Baja Peninsula, as an oceanic arc accreted near 110 Ma (Sedlock, 2003; Wetmore et al., 2003).

The Guerrero terrane is a large composite complex comprising several crustal blocks. This composite block in general represents a juvenile terrane related to the western North America Cordillera. According to a recent compilation (see Centeno García et al., 2008; Centeno-García, 2017 and references therein), this composite block consists of the following terranes: 1) Tahue terrane, which contains Ordovician and Carboniferous-Permian sedimentary rocks related to a continental margin and overlaid with an Early Cretaceous marine arc succession; 2) Zihuatanejo terrane, which is the largest of the Guerrero terrane and

is composed of Early Triassic turbidites with blocks of plagiogranites with MORB signatures; 3) Teloloapan terrane, a Late Jurassic to late Early Cretaceous arc succession; 4) Guanajuato terrane, a late Jurassic to early Cretaceous arc and back-arc succession; 5) the Arcelia terrane, which contains Aptian-Cenomanian mafic and ultramafic rocks, basaltic pillows lavas and hyaloclastites covered by shales, cherts and volcaniclastic turbidites. This composite terrane was accreted to mainland Mexico during the latest Early Cretaceous, and it has been exposed since the Late Cretaceous when it started to supply the foreland basin to the east (Pindell and Kennan, 2009).

The Oaxaquia terrane (Fig. 2), also called microcontinent, is the largest lithospheric fragment in Mexico and is characterized by Grenville-age granulite facies rocks (Ortega-Gutierrez et al., 1995; Solari et al., 2003; 2004). Mesoproterozoic and Neoproterozoic gneisses and anorthosites (e.g., Solari et al., 2003) are covered unconformably by Paleozoic marine to continental sedimentary successions whose fauna indicate a Gondwanan affiliation, and a Middle or Late Paleozoic accretion to North America (Ortega-Gutiérrez et al., 1995).

The basement rocks of the Maya terrane (Fig. 2) include Ediacaran to Permian-Triassic igneous and metamorphic rocks located east of Oaxaquia (Campa and Coney, 1983). This terrane has been interpreted as part of a Late Paleozoic collisional belt associated with the assembling of Pangaea, the Ouachita-Alleghanian-Variscan orogen (Centeno-García, 2005; Martens et al., 2010; Keppie et al., 2011; Ramos-Arias et al., 2015). The Maya terrane can be divided in three provinces: the Yucatan platform province, the northern province including southern Tamaulipas and Veracruz, and the southern province, which includes Chiapas and central Guatemala (Sedlock et al., 1993). The basement of the northern province (Teziutlán Massif; Puebla) is characterized by Grevillian-type rocks, low-grade metamorphic rocks with zircon ages between 330 and 296 Ma, Mesozoic sandstones with detrital zircon populations of 255 and 195 Ma, and Jurassic syenitic rocks with an age of ca. 168 Ma

(Ramos-Arias et al., 2015). The 255 Ma detrital zircons are interpreted as derived from the Permo-Triassic granitoids along the GoM coast.

The Yucatan platform was the locus of the Chicxulub impact. The U-Pb age of zircons extracted from the ejecta (Fig. 3) reflected a primary source with an age of 545 ± 5 Ma (Krogh et al., 1993). Furthermore, there is evidence of Ediacaran granodiorite (546 Ma) from a basement clast in the meteorite ejecta in the Yucatán platform (Keppie et al., 2011). The Maya Mountains in the southeastern margin of the Yucatán peninsula (Fig. 3) are mainly composed of early and late Paleozoic sedimentary rocks, Silurian-Devonian magmatic intrusions and coeval volcaniclastic rocks, which overlie unconformably low-grade metamorphic rocks deposited in the early Paleozoic. The provenance of the early Paleozoic rocks of the Maya Mountains is exclusively from the Grenvillian orogenic belt, which contrasts with the Pan-African and Trans-Amazonian signatures in late Paleozoic rocks of the southern region of the Maya terrane (Martens et al., 2010).

The basement of the Mixteca terrane (Fig. 2) is the Acatlán complex (northwestern region of the Oaxaca-Acatlán block), which consists of high- to low-pressure metamorphosed igneous and sedimentary rocks of oceanic floor, arc, trench turbidites, and reworked Proterozoic crust. These rocks are unconformably covered by Mesozoic and Cenozoic sedimentary and volcanic rocks (Ortega-Gutiérrez et al., 1999). According to Talavera-Mendoza et al. (2005), geochronological data from the Acatlán complex reveal a composite geological history that spans from the assembly of Rodinia to the break-up of Pangaea. They documented three major tectonothermal events in the Esperanza granitoids, which they interpret as part of the southern margin of Laurentia: (1) Mesoproterozoic Grenvillian (ca. 1165–1043 Ma), (2) Ordovician (ca. 478–471 Ma), and (3) Late Ordovician-Early Silurian (461–440 Ma), the latter accompanied by the emplacement of blueschists. Yañez et al. (1991) report Devonian granitoids (ca. 374 Ma), which accompanied an eclogitic metamorphic

overprint. The Laurentian units were juxtaposed along with the Gondwanan Oaxaquia block during the Carboniferous when South America and North America were close to each other, and the collision may be associated with the closure of the Rheic Ocean (Talavera-Mendoza et al., 2005).

The Oaxacan complex (east of the Oaxaca-Acatlán block) is the largest inlier of rocks with Grenvillian-age in Mexico. It consists of para- and ortho-gneisses with protolith ages ranging between ca. 1140 to ca. 1000 Ma, which were affected by tectonothermal events at ca. 1100 and ca. 990 Ma (Solari et al., 2003).

The basement of the Cuicateco terrane (Fig. 1), also described as Juárez terrane by Campa and Coney (1983), consists of Proterozoic and Paleozoic rocks, with mylonite zones, that do not show a clear affinity to the Oaxaquia or Maya terranes. They also reported Devonian–Carboniferous metasediments, and a mafic Permian–Triassic magmatic event. Extrusive rocks and granitoids, the ages of which overlap with each other, suggest an Early Cretaceous magmatic event in the Cuicateco terrane (Delgado-Argote et al., 1992; Coombs, 2016; Coombs et al., this volume). The last magmatic activity in the Cuicateco terrane is late Maastrichtian generation of oceanic floor (Pérez-Gutiérrez et al., 2009a). The occurrence of Cretaceous serpentinites and alkali basalts was interpreted by Mendoza-Rosales et al. (2010) as evidence for a rift-related origin for this terrane. Coombs (2016) and Coombs et al (this volume) proposed a back-arc origin for the Cretaceous alkali basalts of the Cuicateco terrane.

The Mixtequita batholith (northwest of the Chiapas-Mixtequita block in Fig. 3) is formed by Permo-Triassic and Jurassic granitoids (Weber et al., 2005; and references therein). The Mixtequita Permo-Triassic rocks share similar age and composition than Chiapas granitoids, however, Jurassic intrusions have not been reported in the Chiapas Massif. The Mixtequita batholith intrudes Proterozoic granulites of the Guichicovi complex with a probable Oaxaquia affinity (Weber et al., 2005). This implies that Oaxaquia and the

southern part of the Maya terrane, i.e., Chiapas-Mixtequita, were joined at least since the Permian.

The Chiapas Massif (southeastern region of the Chiapas-Mixtequita block; Fig. 3) was described by Weber et al. (2007) as part of the southern Maya Block basement. The massif is dominated by Permo-Triassic granitic rocks that are locally foliated, and by a Permian orthogneiss containing inliers of metasedimentary rocks. Weber et al. (2007) suggested that the Chiapas Massif was part of a Permian active continental margin of western Gondwana which underwent medium- to high-grade metamorphism and deformation during the Latest Permian-Early Triassic coeval with the intrusion of syntectonic granitoids. Furthermore, Ordovician magmatism has benn reported in the southeastern region of the Chiapas Massif at ca. 480–449 Ma in the Motozintla plutons (Estrada-Carmona et al., 2012).

The Xolapa terrane (Figs. 2–3), also known as Chatino terrane, is a composite metamorphic-plutonic terrane (e.g., Campa and Coney et al., 1983; Sedlock et al., 1993). The basement of the Xolapa terrane is the so-called Xolapa complex, located at the northwestern region of the terrane. It was defined by Pérez-Gutiérrez et al. (2009b) as those rocks affected by migmatization and ductile deformation, which predate the intrusion and further solid-state shearing of ca. 130 Ma deformed granites in the Acapulco area in southern Mexico. These authors reported an Early Jurassic magmatic event that intruded a pre-Toarcian succession, originating the oldest paragneisses known in this terrane. They interpreted this suite as an Early Jurassic magmatic arc that fringed the Mixteca terrane, and subsequently was deformed and accreted to the southern Mexico terranes. They also reported that an Early Cretaceous high-grade metamorphism and migmatization event affected this complex. Later, high-grade gneisses and migmatites underwent four tectonothermal events at ca. 55, ca. 45, ca. 45–34, and ca. 34–30 Ma (Pérez-Gutiérrez et al., 2009b). For details we suggest to see e.g., Herrmann et al. (1994); Ducea et al. (2004a); (2004b); and Tolson, (2005).

According to Pérez-Gutiérrez et al. (2009b), several lines of evidence indicate a possible correlation between the Xolapa Complex and the northern margin of the Chortís block. Similar Early Cretaceous migmatization age could represent the collision between the Xolapa Complex and the northern margin of the Chortís block marked by the high-pressure metamorphism in the Motagua zone (Fig. 1). The removal of the Chortís block from southwestern Mexico started during the latest Cretaceous (Solari et al., 2007). The migration event took place along with the emplacement of Paleogene subduction-related plutons in southern Mexico and Chortís, which were finally exhumed during Oligocene–Miocene time (e.g., Ratschbacher et al., 2009; Pindell and Villagómez, this volume, Chapter 14).

The Chortís block is defined as a lithospheric block limited on its northern boundary by the active Motagua zone, and on the south by the Hess escarpment (Dengo, 1969, *in* Torres-de-León et al., 2012). This continental block is considered as an assemblage of heterogeneous terranes and geological provinces whose boundaries and geological evolution remain poorly understood (e.g., Ortega-Gutiérrez et al., 2007; Rogers et al., 2007). The Chortís block was possibly the southern part of the southern terranes of Mexico, and the northwestern part of the Tahamí terrane in northwestern Colombia during the Early Jurassic, according to the model of Pindell and Kennan (2009). They propose that Chortís was detached, and rifted away from the Tahamí terrane (South America) by the Late Jurassic while remaining attached to the southern Mexico lithosphere. During the Early Cretaceous, active-volcanic-arc terranes (such as terranes in Nicaragua, Jamaica and Cuba, the so-called Caribbean arc) laterally accreted by transform faults to the south of Chortís. Due to the evolution of the Caribbean plate, the Caribbean arc migrated northeastwards during the Late Cretaceous while Chortís started to detach from the southern Mexican terranes with an eastward motion by sinistral strike-slip faults (the Motagua Zone in Fig. 3), and continue this motion until its present day location. For a detailed description of the Chortís block evolution

we suggest to see e.g., Morán-Centeno et al. (2009); Pindell and Kennan (2009), Ratschbacher et al. (2009),Torres-de-León et al. (2012), and Pindell and Villagómez, this volume).

The formation of the Sierra Madre Oriental fold-and-thrust belt (SMO), including the Sierra de Zongolica, occurred as part of the Laramide orogeny during the Late Cretaceous and Paleogene, when the post-Jurassic sedimentary and volcaniclastic succession was deformed due detachment-related thin-skin tectonics (e.g., de Cserna, 1989). The uplifted material, Mesozoic carbonates, clastic and volcaniclastic rocks became a source of sediments to the Gulf of Mexico basin. One remaining question is whether this mountain belt barred sediments from the southwestern region (i.e., Guerrero, Oaxaca-Acatlán, Cuicateco-Juárez, and Xolapa) from reaching the Gulf of Mexico during the Oligocene–Miocene. Later, during the mid- to late Miocene, the Chiapas fold-and-thrust belt (Fig. 3) was generated by a compressive-transpressive tectonic episode, the Chiapanecan orogeny, which uplifted sedimentary rocks that became sources of sediment to the Gulf of Mexico basin (Villagómez and Pindell, this volume, Chapter 17). Another open question is whether this orogen obstructed the sediment path from the southern regions such as Chiapas and Mixtequita Massifs, Chortís, and the Central America Arc.

The Trans Mexican Volcanic Belt (TMVB) is the current expression of a long-lasting volcanic arc that has been active since the Jurassic due to the eastward subduction of the Farallón plate and later the Cocos plate beneath western Mexico (e.g., Ferrari et al., 2012). The Jurassic-Cenozoic evolution of the subduction originated an array of arcs, with heterogeneous and dissimilar compositions that partly overlapped. The TMVB adjusted its strike from a NNW-SSE direction (Paleocene) to its current E-W strike (this trend has prevailed since the late Miocene). It is composed of an early Miocene slab-flattening-related andesitic phase, followed by a mafic episode in the mid-late Miocene due to lateral

propagation of a slab tear, and the migration of the volcanic front from west to east, reaching the GoM. In the late Miocene, the volcanic front migrated trenchward, with silicic volcanism, emplacement of intraplate basalts, and extensional tectonics, as the result of slab-rollback in the subduction system. The current arc generates more differentiated rocks or mafic lavas with a mantle fingerprint (Ferrari et al., 2012).

The Los Tuxtlas volcanic field (Fig. 3) is an isolated basaltic complex with alkaline composition in the southern margin of the GoM. Ferrari et al. (2012) suggest that a pulse of mafic volcanism migrated from west to east, reaching the GoM by the late Miocene (ca.7 Ma), which coincides with the age of the onset of volcanism in the Los Tuxtlas volcanic field. This suggests that the origin of this volcanic field may be related to the evolution of the TMVB.

A Cenozoic paleo-canyon network located along the western margin of the GoM (Fig. 3) possibly delivered sediments to the southwestern region of the GoM basin from the erosion of Paleocene and Mesozoic rocks during the Paleogene uplift of the SMO. Those deep-incised paleochannels developed on the foreland basin of the SMO orogen under submarine conditions, and its filling material is characterized by turbidite facies. The material deposited in this paleochannels has economic relevance, inasmuch as these successions enclose hydrocarbons. The Chicontepec and the Bejuco-La Laja paleocanyons are elongated basins rather parallel to the SMO orogen. The Bejuco-La Laja paleocanyon is filled by fragments of Jurassic-Cretaceous calcareous rocks and Paleocene shales in a neritic to bathyal marine environment (Cantú-Chapa, 1987; Alegría-Montenegro, 2015).

Vásquez et al. (2014) proposed a stratigraphic evolution of the Chicontepec paleo-canyon characterized by the overlapping of submarine fans with two contrasting paleo-flow directions: i) during the early Paleocene the submarine fans indicate a NW-SE flow direction, which is parallel to the SMO orogen, and the provenance could be affiliated to a northern

region of the foreland basin system of the SMO, such as La Popa and/or Parras basins, and ii) mid to late Paleocene submarine fans indicating SW-NE flow direction, which is perpendicular to the SMO, with a provenance associated to the SMO. Recent works compiled in Cossey et al. (2016) suggest that the paleocanyons were eroded by fluvial systems feeding directly into the central GoM basin, probably a land-locked sea during the Paleocene-Eocene boundary. That study interpreted dramatic changes of the sea level as the result of a drawdown episode of the GoM (including Chicontepec paleo-canyon) due to the isolation of the GoM from the world's oceans, as the result of the closure of the Florida Straits as the Cuban arc collided with the Bahamas and northeast Yucatán. This interpretation is in line with the model of the Paleogene isolation of the GoM proposed in Rosenfeld and Pindell (2003). The timing of the interpreted drawdown coincides with the Paleocene-Eocene Thermal Maximum (e.g., Röhl et al., 2000), hinting that this thermal event may have been caused or assisted by the release of methane from hydrates in the GoM margins and abyssal plain.

2: METHODS

Sample preparation method

Samples were treated with hydrogen peroxide 10% and acetic acid 11%, and went through ultrasonic bath to liberate the grains from matrix/cement remnants. Then, the samples were sieved and sorted into 32–63 μ (coarse silt), and 63–400 μ (very fine sand to medium sand) fractions. Optical observations indicate higher presence of barite in the silt-size fraction; thus, analyses of the fine-sand fraction were preferred. Mineral separation was carried out following the traditional heavy liquid methodology (Mange and Maurer, 1992) in order to separate heavy and light minerals. The samples from Puskon-1 and Aktum-1 sections were analyzed by the SEM method, and the rest of the samples were analyzed by their XRD patterns.

SEM analysis: heavy and light minerals

Heavy and light mineral fractions were mounted in 2.5 cm diameter epoxy resin mounts. The mounts were polished with diamond paste to expose the minerals and subsequently coated with carbon. Analyses were carried out in a JEOL JSM-6390 LA instrument, equipped with a back-scattered electron (BSE) detector and an energy-dispersive spectroscopy (EDS) analyzer (EDAX) at the Institute of Geochemistry and Petrology, ETH Zurich. The tungsten filament of the SEM was operated with an acceleration voltage of 15–17 kV, a filament current of typically 50–70 μA, and the sample was placed at a working distance of 10–14 mm and time acquisition was 30 s. The NSS Spectral Imaging System software of Thermo Fischer Scientific Inc. was used to automatically collect ED spectra and semiquantitative elemental concentrations. Spectrum files can be reprocessed to include accidentally omitted elements retrospectively, without the need to physically re-analyze the sample.

The mineral classification is purely based on the chemical composition of the measured grains. No further mineralogical or crystallographic investigations are involved in the classification, e.g., the TiO_2 (i.e., rutile, anatase and brookite) and Al_2SiO_5 (andalusite, sillimanite and kyanite) polymorphs cannot be distinguished, and, therefore, they are grouped as macro-chemical mineral families (Commeau et al., 1992). The distinction between ilmenite, leucoxene and rutile, for example, is entirely based on the measured TiO2 wt% content only.

The amount of water in the crystal lattice cannot be determined with EDS analyses for two reasons: 1) it is not possible to measure light elements, like B, C, or O, precisely, and very light elements, like H, at all, and thus, it cannot be determined whether water is present or not; 2) the EDX software automatically recalculates the obtained concentrations to 100%, since analyses are based on standard-less analysis. Therefore, it is not possible to evaluate if

certain elements are not represented in the analysis. As a result it is not possible to distinguish between pyroxenes and hornblende. A large compositional overlap exists between these two mineral groups if water in the hornblende crystal lattice is not taken into account. Therefore, the mineral classification scheme lists both minerals together: amphibole and pyroxene. For the same reasons, no distinction can be made between magnetite and hematite. Distinctions between amphiboles and pyroxenes were made by optical interpretations of the author. Between 500 and 800 grains were individually analyzed in each of the samples of Puskon and Aktum sections.

XRD analysis: heavy and light minerals

The heavy and light mineral fractions were divided by their magnetic susceptibility in a Frantz magnetic separator at various amperages (amp). The obtained fractions were: 1) extreme magnetics (separated with a common magnet), 2) magnetics at 0.4 A, 3) magnetics between 0.4 and 1 0.0 A, and 5) higher than 1 0.0 A. All magnetic fractions were weighted in order to know their proportions within a single sample. Each magnetic sample was crushed in an opal mortar-and-pestle, and subsequently mounted in a flat watch-shaped glass.

The XRD analysis were performed in a latest generation Powder X-ray Diffractometer (Bruker, AXS D8 Advance), equipped with a Lynxeye superspeed detector and an automatic sample changer for 9 specimens. The diffractometer was controlled by the DIFFRAC plus XRD software and measurements were 60–90 min per each analyzed magnetic fraction. We used EVA software to discriminate the mineral phases present in each fraction by a comparison with an internationally recognized database of reference XRD patterns for more than 70.000 phases, and to calculate the concentration level of each recognized phase. This automated computer-controlled diffractometer system allowed us to measure, record and subsequently interpret the unique diffractograms produced by the individual constituents in complex mixtures such as sedimentary deposits.

LA-ICP-MS U-Pb dating of detrital zircons

Detrital zircons were handpicked from the heavy mineral suite of each analyzed sample. Due to the small amount of material per sample we merged the heavy mineral suite of some adjacent levels in order to reach a statistically valid number of grains for a valid analysis. This approach allowed us to present the analysis of detrital zircons from 12 sandy levels.

Picked zircons were mounted in epoxy and polished to expose their half sections. Then, they were coated with carbon and their internal structure was analyzed by cathodoluminiscence imagery (CL) in order to evaluate magmatic growth zonation, metamorphic origin, or inherited cores. For this purpose we used the Vega 3W-SEM at the center of electronic microscopy ScopeM at ETH Zurich.

In situ U-Pb geochronology of detrital zircons were conducted by LA-ICP-MS at the Institute of Geochemistry and Petrology, ETH Zurich using a 193 nm Resolution (S155) ArF excimer laser (30µm laser diameter) coupled to an Element SF ICPMS (Guillong et al., 2014; von Quadt et al., 2016). The output energy was typically ca. 2 J/cm^2 and a 5 Hz pulse repetition rate was used. The ablation was under helium flow of 0.7 L/min. Argon was admixed to the aerosol within the funnel of the ablation cell to transport the ablated material to the ICP for ionization. Dwell times range from 5 – 30 ms and peak hopping was employed. Oxide generation was optimized at ThO^+/Th^+ = <0.3%. For each analysis a baseline was measured for 30 seconds followed by 30 seconds of ablation. Elemental concentrations were calculated using the IGOR based Iolite software (Paton et al., 2011). The stoichiometric Si concentration of 15.2 wt.% for zircon was used as an internal standard. SRM NIST 612 glass (Hinton, 1999) was used as the primary external trace element standard and was measured four times each 25 zircon analyses.

The masses 202, 204, 206, 207, 208, 232, 235 and 238 were measured. Total ablation time was set to 30 seconds with a gas blank/background measurement of 10s. Age data were collected in runs of 20 samples bracketed before and after by two analyses of the primary reference material GJ-1 (Jackson et al., 2004) and each one of secondary reference zircons 91500 (Wiedenbeck et al., 1995), Temora (Black et al., 2003; von Quadt et al., 2016), Plesovice (Sláma et al., 2008), AusZ7.5 (von Quadt et al., 2016) and SA2 (von Quadt, not published, personal commun.). Data reduction was performed with the IGOR based Iolite v2.5 (Paton et al., 2011) and Vizual Age (Petrus and Kamber, 2012) software. Obtained isotope ratios and dates are corrected for mass bias, instrumental drift and downhole fractionation using primary reference material. Analyzed ages and trace elements of spots with elevated Al, P, Ca, Mn or Fe were discarded as these values indicate the presence of mineral (e.g., apatite, feldspar) or melt inclusions.

Statistical analyses of zircon data were performed using Isoplot 3.75 and 4.15 (Ludwig, 2012). All discordant (> 1%–3%) analyses of magmatic zircons were discarded. Only zircons with concordance greater than 90% were accepted and plotted.