Permian to quaternary magmatism beneath the Mt Carmel area, Israel: Zircons from volcanic rocks and associated alluvial deposits

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A B S T R A C T

Xenocrystic zircons from Cretaceous pyroclastic vents on Mt. Carmel, N. Israel, document two major periods of earlier mafic magmatism: Permo-Triassic (285–220 Ma) and Jurassic (200–160 Ma). Related alluvial deposits also contain these zircon populations. However, most alluvial zircons are Cretaceous (118–80 Ma) or younger, derived from Miocene to Pliocene volcanic episodes. The Permo-Triassic-Jurassic zircons are typically large and glassy; they have irregular shapes and a wide variety of internal zoning patterns. They appear to have grown in the interstitial spaces of coarse-grained rocks; many show evidence of recrystallization, including brecciation and rehealing by chemically similar zircon. Grains with relict igneous zoning have mantle-like 6°/3° (5.5 ± 1.0%), but brecciation leads to lower values (mean 4.8%, down to 3.1%), Hf-isotope compositions lie midway between the Chondritic Uniform Reservoir (CHUR) and Depleted Mantle (DM) reservoirs; Hf model ages suggest that the source region separated from DM in Neoproterozoic time (1500–1000 Ma). Most Cretaceous zircons have 18°/17°/175°/176° similar to those of the older zircons, suggesting recrystallization and/or Pb loss from older zircons in the Cretaceous thermal event. The Permo-Jurassic zircons show trace-element characteristics similar to those crystallized from plume-related magmas (Iceland, Hawaii). Calculated melts in equilibrium with them are characterized by strong depletion in LREE and P, large positive Ce anomalies, variable Ti anomalies, and high and variable Nb, Ta, Th and U, consistent with the fractionation of monazite, zircon, apatite and Ti-bearing phases. We suggest that these coarse-grained zircons crystallized from late differentiates of mafic magmas, ponded near the crust-mantle boundary (ca 30 km depth), and were reworked repeatedly by successively younger igneous/metamictic fluids.

The zircon data support a published model that locates a fossil Neoproterozoic plume head beneath much of the Arabia-Levant region, which has been intermittently melted to generate the volcanic rocks of the region. The Cretaceous magmas carry mantle xenoliths derived from depths up to 90 km, providing a minimum depth for the possible plume head. Post-Cretaceous magmatism, as recorded in detrital zircons, shows distinct peaks at 30 Ma, 13 Ma, 11.4 ± 0.1 Ma (a major peak; n = 15), 9–10 Ma and 4 Ma, representing the Lower and Cover Basalts in the area. Some of these younger magmas tapped the same mantle source as the Permain-Jurassic magmatism, but many young zircons have Hf-isotope compositions extending up to DM values, suggesting derivation of magmas from deeper, more juvenile sources.

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1. Introduction

Zircon is an excellent tracer of the timing and nature of magmatic activity. It also can be used when magmas remain in the upper mantle or deep crust, where they can be sampled by later eruptive magmas. This approach is directly analogous to detrital-zircon analysis in crustal settings, where zircons are used to investigate the magmatic and metamorphic history of rocks in the drainage area. As with detrital-zircon studies, this tool becomes especially effective when U—Pb dating is combined with the analysis of Hf—O isotopes and trace elements.

Several studies of kimberlites (Belousova et al., 2001; Kinny et al., 1989; Tretiakova et al., 2017) have identified multiple populations of mantle-derived zircons, demonstrating that zircon can retain crystallization ages despite high temperatures; it thus can record magmatic episodes for which surface expression has been lost, or not recognized. This is less commonly possible with more mafic types of volcanic rocks, perhaps because xenocrystic zircons have been dissolved or perhaps because the presence of magmatic zircons is not expected. In general, zircon does not crystallize directly from mafic magmas, because of the high solubility of Zr in such melts. However, if mafic magmas stall...
and fractionate during long slow cooling in the upper mantle or lower crust, zircon can crystallize from late residual melts or fluids. In areas of continuing magmatic activity, such zircons can be entrained in the next erupting magma, and their ages may be indistinguishable (within errors) from the age of the host basalt. Alternatively, the deep-seated zircons may react with later melts and fluids to give complex age patterns, when they are sampled by still later magmas. The analysis of zircon xenocrysts in mafic rocks thus can be used to examine the magmatic history of the uppermost mantle and lower crust. In this paper we use this analytical tool to investigate the history of basaltic magmatism, as sampled by late Cretaceous volcanoes on Mt. Carmel, northern Israel, with special emphasis on events of crustal underplating from the breakup of Gondwana to the Pliocene. A secondary aim has been to constrain the ages of some of the eruptive centers and to understand differences in the sampling patterns of individual centers, as a guide to exploration.

Late Cretaceous magmatism in the Mt. Carmel area (Fig. 1) involved repeated pyroclastic eruptions in a shallow marine-shelf setting, building several km-scale seamounts over the period from ca 97–80 Ma (Segev and Sass, 2009). Although no older volcanic rocks outcrop within the region, the Cretaceous vents and tuffs described here contain several populations of older zircons with distinctive characteristics, revealing a history of mantle-derived magmatism stretching back to Lower Permian time (ca 280 Ma). These occurrences offer new insights into magmatism related to the breakup of Gondwana, and raise questions surrounding the behaviour of the lithospheric and asthenospheric mantle during rifting, and the conditions that would focus repeated mafic magmatism into this small area (ca 150 km²) since Permian time.

Here we present U—Pb dating, O—Hf isotopic analyses and trace-element data from ~400 zircons, collected both from the Cretaceous pyroclastic rocks and from associated alluvial deposits. We use these data to constrain the ages of some of the Cretaceous eruptive centers, to map out the mantle distribution of intrusive magmas related to the older volcanic episodes, and to discuss the questions flagged above.

1.1. Regional setting

The material described here is derived from Cretaceous volcanic centers and related alluvial deposits in the Mt. Carmel-Yizre’el Valley area of northern Israel (Fig. 2). This area lies within a complex system of minor rifts and other faults, related to the Africa-Arabia plate boundary that later developed into the Dead Sea Transform, with ~100 km movement since the initiation of offset in Miocene time. The NW-SE Carmel Fault that bounds Mount Carmel on the east (Fig. 1) is part of the Carmel-Gilboa system, which forms a 10–20 km wide belt of faulting running through the Yisre’el Valley, and may extend across the continental margin (Segev and Rybakov, 2011). The SW side of Mount Carmel is dissected by multiple N-striking vertical faults with offsets up to a few hundred meters. The basement rocks, which are not exposed in this area, are considered to have formed in Pan-African (Cadomian) time (>620 Ma; Stein and Goldstein, 1996). The Galilee area represents a zone of thin continental crust, with a Moho depth of 23–32 km (Segev and Rybakov, 2011). The geophysical data, and estimates of the Cretaceous geotherm from mantle-derived xenoliths and xenocrysts (Apter, 2014; our unpublished data) suggest a thin (<100 km), hot lithospheric mantle. It is not clear whether this is a remnant of a previously thicker lithosphere that was thinned during rifting and drifting, or has developed later through magmatic processes (Stein and Hofmann, 1992).

This region experienced considerable mafic and felsic volcanism during and following the Permo-Triassic Palmyra rifting, related to the breakup of Gondwana and the opening of the Neo-Tethys ocean refs. The northern end of Mt. Carmel is the centre of a major NW–SE magnetic anomaly, which extends well offshore. Modelling by Segev and Rybakov (2011) indicates that the magnetic source is buried at least 5 km below the surface, and the authors suggest that it reflects magmatism related to the Palmyra rifting. This was followed by a major Early Jurassic (197–193 Ma) event which erupted up to 2.4 km³ of alkali-olivine basalts and pyroclastic rocks with OIB-intraplate geochemical signatures (Dvorkin and Kohn, 1989). A high positive gravity anomaly centered...
under the northern end of Mount Carmel has been interpreted to reflect the existence of a buried shield volcano beneath Mt. Carmel (Ben-Avraham and Hall, 1977; Gvirtzman et al., 1990). However, a more recent study (Segev and Rybakov, 2011) suggests that the anomaly is only high relative to deep basins on either side, which are filled with thick clastic sediments.

The early development of the late Jurassic-Early Cretaceous Levant magmatic province, extending ca 800 km from central Syria to the Gulf of Suez, coincided with major rifting across Africa, but no major rifting events appear to correlate with the relatively sparse distribution of Early Cretaceous (C1, 137–139 Ma; Segev, 2005; references therein) continental magmatism within Israel. This magmatism comprises subalkaline to alkaline basalts, nephelinites, basanites and microgabbros with hotspot/mantle plume geochemical and isotopic signatures (Garfunkel, 1989; Laws and Wilson, 1997; Stein and Hofmann, 1992; Stein and Hofmann, 1994). This type of magmatism...
continued until ca 125 Ma in Lebanon and possibly into northern Israel (C2; Segev, 2005).

The Late Cretaceous (98–94 Ma, Turonian–Cenomanian) volcanic activity of northern Israel took place in the Mount Carmel–Umm El Fahm area (Fig. 2), and has been described by Sass (1980). The dominant volcanic rocks are pyroclastics of maﬁc to ultramaﬁc composition; lava ﬂows played a minor role in this volcanic activity. The volcanism took place in a shallow marine environment, in intimate contact with marine sediments containing rich assemblages of fossils. Several outcrops of black and variiegated pyroclastics in isolated areas west of Kerem Maharal (Fig. 2) overlie the Yagur Formation of Albian (Bein, 1974; Rosenfeld and Raab, 1984) to earliest Cenomanian age (Lewy, 1991) and represent the ﬁrst volcanic eruptions in the area. The black pyroclastics appear to represent eruptive vents, while the variiegated pyroclastics occur as layers of various thickness interbedded with the carbonates, indicating repeated explosive eruptions and the construction and levelling of small seaamounts. The eruptions continued up to ca 94 Ma, represented by the Shefaya volcano (Fig. 2).

There is little published information on the composition of the Late Cretaceous eruptive products, which are variably altered by contact with sea water. Area-scan EDS analyses of ﬁnely-crystalline lapilli from the Rakefet Magmatic Complex have compositions corresponding to tholeiitic picrites, but even these apparently fresh samples may have lost alkali elements during diagenesis. The pyroclastic deposits carry mantle-derived xenoliths (spinel peridotites, garnet + spinel pyroxenites, rare garnet websterites), crustal xenoliths ranging from garnet-ilmenite granulites to Triassic limestones, megacrystic Mg-ilmenite, amphibole and clinopyroxene, and xenocrysts (olivine, orthopyroxene, diopside, pyrope garnet) derived from mantle peridotites, pyroxenites and eclogites. P-T estimates on rare garnet websterite xenoliths indicate they were derived from depths up to ca 90 km (Esperança and Garfunkel, 1986; Mittlefehldt, 1986; Apter, 2014; Kaminchik et al., 2014; our unpublished data), giving a minimum depth for generation of the host magmas.

Abundant Neogene volcanic rocks extend throughout the region both on the surface and in the subsurface, and are divided into two main volcanic episodes. The earlier episode (Lower Basalt Fm.) is assigned to Mid-Late Miocene and was characterized by extension and faulting across the entire region, coinciding with the initiation of lateral-lateral movement on the Dead Sea Transform. This extensional stress regime caused the break-up of the crust, including reactivation of NW-SE fault systems and graben development, opening paths for volcanic eruptions focused on the valley margins, adjacent to normal faults. The younger volcanic period (Cover Basalt Fm.) is assigned to the Pliocene and characterized by further faulting of the valleys. Large-scale volcanism prevailed across Saudi Arabia, Jordan, Syria and northern Israel (Segev, 2005), and large areas of central and eastern Galilee as well as the Golan Heights were covered by basalt ﬂows (Weinstein et al., 2006), originally totaling ca 300,000 km³ (Shaliv, 1991). Later faulting and accumulations of the Cover Basalt in the range of 0–150 m, and minor eruptions between these two episodes, are expressed in outcrops, mostly in eastern Galilee. In the lower Miocene, between the eruptive periods of the Lower and Cover basalts, the area was inundated by a marine transgression, which had important implications for the distribution of the heavy minerals eroded from the Cretaceous volcanics. For example, beach placers (quartzites with carbonate cement) containing abundant moissanite, garnet and spinel have been sampled near Migdal-el-Haq during the Shefa Yamim exploration program, far from known Cretaceous volcanic centres.

1.2. Local setting

The late Cretaceous volcanism on Mount Carmel has been described in elegant detail by Sass (1980). Despite widespread normal faulting, he was able to identify 16 volcanic events (Table 1); in 9 cases the corresponding vent was identiﬁed, while in three other cases the vents are buried under younger sediments. Lavas play a minor role in this volcanism; four sources were identiﬁed but none is correlated with the known explosion vents. The typical Mount Carmel volcano was a pyroclastic cone, erupted in shallow water (<60 m deep), and later eroded to below the water level. Vents are up to several hundred meters across, and are characterized by dense black pyroclastics, while the ﬂanks consist of variably altered air-fall and water-laid pyroclastic debris, dipping away from the cone. Estimated heights of the uneroded cones are from 100 to 350 m, and the ash-fall deposits extend for several kilometers around some of the vents, where not disturbed by faulting (e.g. Makura, Fig. 2).

The most complicated volcano described by Sass (1980) is the Rakefet Magmatic Complex, the outcrops of which extend for ca 2.5 km and are bounded by faults on either side. The Complex comprises three separate eruptive events. The vent of the ﬁrst Rakefet volcano is marked by black pyroclastics exposed over at least 60 m vertically. Our own ﬁeld investigations suggest that at least three eruptions occurred within the area of the main vent. The bedded pyroclastics of this vent are overlain by those of the second volcano, from a vent south of the exposed area. The third volcanic event comprises ca 25 m of olivine basalt and andesine olivine basalt, some vesicular. Their source is not located but is presumed to be nearby. The Rakefet Volcanic Complex is the only example in the area of repeated magmatism within a small area. The Muhraka volcano to the north, known mainly from bedded pyroclastics, may be related to this complex (Fig. 2).

Mount Carmel is bounded on the south by the sediments of the Ramot Menashe syncline (Fig. 1). Drilling (Ein Ha-Shofet #1 deep monitoring drill, Mekorot Ltd) on a magnetic high near the axis of the syncline intersected volcanic rocks at 788–795 m depth, near the base of the Deir Hanna formation. The rocks at this level are tuffaceous, and some horizons are marked by abundant large euhehedral, zoned crystals of high-Mg calcite. Several other basalt/tuff horizons were intersected higher in the Deir Hanna formation (550–750 m). The Uhm-al-Fahn area, still further south of Mt. Carmel (Fig. 2), exposes a thick section of basalts, which have not been sampled in this study.

The Shefa Yamim exploration program has sampled many of the smaller streams and rivers in the drainage basin of the Kishon river, lying to the east and southeast of Mount Carmel (Fig. 2). Bulk sampling has been focused on the mid-reach area, where the Kishon River ﬂows through a narrow gap between Mount Carmel and the adjoining uplands, creating a natural trap for transient placers. Most of the samples labelled “alluvial” in this paper are from this area.

1.3. Sampling

The Shefa Yamim exploration project is aimed at the discovery of economically viable placer deposits of gemstones (mainly sapphire, ruby, hibonite, moissanite) and other commodities within the drainage basin of the Kishon River. The project has sampled the Cretaceous pyroclastic centers on Mt. Carmel (Fig. 2; Rakefet, Har Alon, Bat Shelomo, Muhraka and Beit Oren complexes) and adjacent areas (Ein Ha-Shofet), and minor and major drainages in the Yaz’aerial Valley. Samples range in size from several kg to >1000 t. All samples were screened through a static grizzly screen to remove pieces larger than 100 mm in diameter. Rock samples from the vents were coarsely crushed and then treated in the same way as alluvial samples. The <100 mm fraction was washed in a scrubber that breaks up any clods. The <0.5 mm component is suspended in the wash water and pumped to settling ponds; fractions larger than 25 mm are used to backﬁll exploration pits. Samples in the +8 mm–16 mm and +16 mm–24 mm size fractions are sorted by hand on a picking belt. The +0.5–8 mm component of the sample is washed and classiﬁed into 5 fractions: 0.5–0.7 mm, 0.7–1 mm, 1–2 mm, 2–4 mm, 4–6 mm, 6–8 mm. These fractions are transferred to a pulsating jig plant for gravity separation. Samples in the 2 mm–8 mm size fractions are visually inspected after the jigging process and sorted in the recovery laboratory. The three smallest size fractions are jigged...
separately; most zircons in this study come from these fractions. The heavy concentrate in the center of the jig pan is collected and dried; material on the outer part of the jig pan is discarded. The sorters in the laboratory have demonstrated their efficiency in identifying and recovering a wide range of mineral species, including garnet (pyrope), ilmenite, spinel, chrome-diopside, diamond, moissanite, sapphire, ruby, Carmel Sapphire™ and hibonite, as well as rutile and zircon. The sample set described here is dominated by zircons from the Rakefet Magmatic Complex, and the alluvial deposits of the Kishon River mid-reach, because these localities have been bulk-sampled, providing abundant material.

The unprocessed heavy mineral concentrates of several samples were hand-picked under a binocular microscope in the CCFS laboratories; a few rock samples also have been processed by SelFrag (electrostatic disaggregation) techniques at CCFS, sieved and hand-picked after magnetic and heavy-liquid separation.

1.4. Sample descriptions

Two general groups of zircons in this study can be distinguished by combinations of external morphology and internal structure, here designated “Mount Carmel zircon” (MCZ) and “common zircon” (CZ).

MCZ is the most distinctive group. It comprises generally large grains (0.5–2 mm) with blocky to rounded to irregular shapes, including pronounced rounded knobs and depressions (Fig. 3). Prism or pyramid faces are visible on some grains but usually show rounding and resorption. These grains are typically colourless and clear, with a brilliant luster. Most MCZ have low Cl response, but images of some grains show patterns that suggest relict, often blurred, igneous or metamorphic zoning, including core–rim structures with irregular cores. Many grains appear to be fractured or even brecciated (see below). Similar irregular zircons have been separated from kimberlite-borne peridotite xenoliths (A. Guliani, pers. comm. 2016), suggesting that this morphology was imposed when the zircons crystallized from melts in the interstitial spaces between other minerals, either in cumulates or in the peridotitic wall-rock. The outer surfaces of some grains show an unusual ‘bubble’ texture, suggesting partial melting of the grains, perhaps during magmatic transport.

For the purposes of this paper, the MCZ zircons can be subdivided into three classes: igneous, igneous/metamorphic and breccia grains. “Igneous” grains show oscillatory zoning, or broad lamellar zoning, regardless of external morphology; however, most have more regular shapes than other MCZ grains. Some grains have distinct cores (usually metamict, some with contorted zoning) surrounded by thick homogeneous rims with lighter or darker CL response (Fig. 3a). While the grains with homogeneous CL could be regarded as metamorphic, the continuous transition toward the “igneous” grains, in terms of internal structure and external morphology,
persuades us that most had an igneous heritage. None of these grains has the low Th/U (<0.1) commonly regarded as indicative of metamorphic zircon, though this criterion may not be applicable to zircons in mafic or ultramafic rocks. These issues are discussed further below in the light of the isotopic data.

“Breccia” grains generally have low CL response and no traces of zoning, and are typically irregular in shape. They appear to have been fractured and rehealed. The differences in CL suggest that the internal fragments are slightly misaligned, but the fractures are sealed with zircon, in optical continuity with the host grain. In some grains brecciated cores are overgrown by featureless rims (Fig. 3d). Zircons with similar brecciated/rehealed internal structures have been found in heavy-mineral separates from Siberian kimberlites (Tretiakova et al., 2017).

CZ from the vents and the alluvial deposits display a wide range of morphology, ranging from prismatic to rounded. Oscillatory to lamellar zoning in CL images indicates that most of the CZ grains are igneous in origin, but some show evidence of metamorphism. The alluvial deposits contain a higher proportion of broken and/or rounded grains, as might be expected. The MCZ population is relatively abundant in the vent tuffs, but much less common in the alluvial deposits, suggesting that the brecciated grains, in particular, may not withstand transport as well as the CZ population.

1.5. Analytical methods

The zircons have been characterized using optical microscopy, scanning electron microscopy (SEM) for CL imaging, electron microprobe (EMP) for major-element analysis, laser-ablation (LA)-ICPMS for U—Pb and trace-element analyses, secondary ion mass spectrometry (SIMS) for U—Pb and O-isotope analysis, and LA-multicollector (MC)-ICPMS for analysis of Hf isotopes. Brief details of these methods are given in the Appendix. The zircons have been classified in terms of their primary host rock using their trace-element chemistry and the classification schemes of Belousova et al. (2001).

2. Results

2.1. U—Pb ages

The U—Pb dataset comprises 240 analyses of zircons from the vents and tuffs, and 190 analyses of detrital zircons. Essentially all analyzed zircons are concordant (Table A1; Fig. 4). In this paper, ages <1 Ga are 206Pb/238U ages; older ones refer to 207Pb/206Pb ages.

The zircons from the vents show age peaks in the Permian (ca. 285–260 Ma), Triassic (ca. 240–220 Ma) and lower Jurassic (ca. 210–185 Ma; correlative with the “Asher volcanics” known from deep wells), which tails off with time to ca. 175 Ma. Bulk sample #479 of the Rakefet Magmatic Complex yielded five older grains with ages of 3124, 2726, 538 and 420 Ma. The Ein Ha-Shofet sample contains more older grains, including two Neoarchean grains (2575, 2635 Ma) and several Neoproterozoic ones (614, 757, 796, 950, 975 Ma). Most of these are clearly igneous, with well-developed CL zoning, but many are well-rounded or fragmental. Some may be detrital zircons, sampled from sediments intercepted at depth by the basalt during eruption. However, the 2767-Ma grain from Rakefet is overgrown by zircon dated at 206 Ma, and presumably was caught up in the Jurassic magmatism.

The individual vents vary widely in their zircon populations (Table 1), although some of this may reflect the volumes sampled from different sites. The zircon suite (n = 100) from the Rakefet Magmatic Complex, which was bulk-sampled, covers the entire range of Permian to
Jurassic ages, but has only one grain relevant to the age of eruption (93 ± 3 Ma). The Bat Shelomo locality is dominated by Permo-Triassic zircons (mean age 253 ± 15 Ma, \( n = 18 \)). It also includes one old grain (283 ± 3 Ma) and one grain (99 ± 4 Ma) that may be close to the age of the vent. However, the sample also contains two Miocene zircons (13–12 Ma) whose origin is unknown; they may be related to the Miocene marine transgression described above. The Beit Oren complex also contains zircons with ages ranging from 286 to 190 Ma, but no younger ones. The Makura complex yielded only 2 grains, with ages of 282 and 281 Ma. The Ein Ha-Shofet sample yielded 76 zircons, of which 50 were dated. Most fall into two clearly separated populations: 190 ± 3 Ma (\( n = 12 \)) and 273 ± 6 Ma (\( n = 22 \)). One grain at 99.9 ± 1.4 Ma constrains the age of the eruption. The Tavassim vent yielded one Permian grain; the remainder belong to the Cretaceous suite, ranging from 128 to 84 Ma (\( n = 4 \)). Similarly, the Karem Maharal tuffs yielded two Permian zircons (276 and 248 Ma) and three Cretaceous ones (107–90 Ma), providing some constraint on the eruption age. Four of the five zircons from Har Alon are
Jurassic (199–193 Ma) but a younger grain (98 Ma) may date the eruption.

The alluvial deposits (Fig. 5c) contain surprisingly few of the pre-Cretaceous zircons (MCZ) that dominate the populations from the sampled vents (Fig. 5b); nearly all of the zircon grains fall into the “common zircon” group described above. However, the zircons from the stream sediments show a major complex age peak extending from 120 to 75 Ma, indicating that Cretaceous vents have in fact been sampled; the scarcity of Jurassic-Permian zircons thus may reflect a lower survival rate of these complex zircons during alluvial transport. The presence of late Cretaceous zircons with ages (75–70 Ma) younger than the sampled vents suggests the (former?) existence of undiscovered vents, and the continuation of the volcanism into Campanian time. The alluvial samples also contain zircons with a scattering of ages (Fig. 5d) extending from the Eocene up into the Pliocene, with peaks at ca 30 Ma, 13 Ma, 11–12 Ma (a major peak; \( n = 15 \), 11.4 ± 0.1 Ma, MSWD = 0.79), 9–10 Ma, and 4 Ma. These appear to represent zircons from the Lower and Cover Basalts.

Brecciated zircons from the Permian-Jurassic populations can show spot-to-spot internal variations ranging from a few tens of Myr, up to 190 Myr. In general, the more homogeneous blocks retain the oldest ages, while boundaries between domains are younger, but several exceptions were noted. 17 core-rim pairs were analyzed, mostly within the 250–200 Ma population. However, one Archean (2767 Ma) core had a distinct 206 Ma rim, suggesting that the Archean grain was picked up by the host magma emplaced in Triassic time. One Jurassic core (188 Ma)
Ma) had a Cretaceous rim (104 Ma), and one late Cretaceous (74 Ma) core had an Oligocene rim (31 Ma). These pairs are discussed further below, with the Hf-isotope data.

2.2. Trace-element compositions

Extended chondrite-normalized REE patterns for 90 zircons from the different vents are gathered in Fig. 6; the analytical data are compiled in Table A2. All but one of the analyzed zircons from Bat Shelomo are classified as igneous. Most of the Permo-Triassic zircons have very similar patterns, with an overall gently positive slope from Sm to Lu, and strong negative Eu anomalies coupled with large positive Ce anomalies. One grain with a metamict core shows disturbance in the form of enrichment in LREE. One Triassic grain (233 Ma) differs significantly in having lower LREE and no Eu anomaly, and this pattern is similar to those in both the Cretaceous and the Miocene zircons. The Miocene grain has much higher REE than the other grains.

A similar pattern is seen at Kareem Maharal, where the two Permian grains have high HREE and negative Eu anomalies, while the Cretaceous grains have flatter patterns with no Eu anomalies. At Tavassim and Beit Oren, this pattern is repeated; Permian grains show an Eu anomaly while the Cretaceous grains do not. The metamict cores of two older grains show an enrichment in LREE. The zircons from Har Alon are Jurassic from one Cretaceous one. The patterns of the Jurassic zircons resemble those from the older populations, but with shallower Eu anomalies; the one Cretaceous grain has a flatter pattern and no Eu anomaly.

The relatively few obviously igneous zircons from the Rakefet Magmatic Complex all have ages in the range 253–175 Ma and have very similar patterns, with modest Eu anomalies in the oldest grains and smaller ones in the younger ones. They show a wide range in Th and U contents, and Th/U > 1. The much larger number of igneous/metamorphic grains shows a wider range in absolute REE contents, and especially a larger spread in the HREE than in the other suites discussed above. Several of the patterns have flat to slightly negative slopes from Dy to Lu, combined with very minor Eu anomalies. The spread in Th and U concentrations covers more than two orders of magnitude. In striking contrast, the brecciated grains are more homogeneous, aside from a range of Eu anomalies.

The variations in REE patterns, U/Th relationships and the Eu anomalies through time are illustrated in Fig. 7. The plot of La/Sm vs age (Fig. 7a) shows that despite general similarities, especially in the slope of the HREE, the Permian to Jurassic zircons tend to have more LREE-enriched patterns than the Cretaceous and younger zircons. The more strongly negative Eu anomalies, and lower Ce anomalies, of the Permian to Jurassic zircons, are obvious in Fig. 7bc. There is no clear relationship between Eu/U* and the morphological classifications. There appears to be an overall decrease in the depth of the Eu anomaly from the Permian to the Jurassic, but it is not clear if this is significant.

The abundances of Ce, U and Th have been converted to estimates of oxygen fugacity (fO2) using the formulation of Loucks and Fiorentini (2018). The estimates (Table A2; Fig. 7d) cover >4 log units, centered around the fO2 of the QFM buffer; there is no clear age trend in the data, but the three 9–12 Ma zircons recorded higher fO2 than most of the older grains.

The Th/U ratios of the zircons show some change with time; Th/U > 1. The more common values >0.3 are much more common in the Miocene to Pliocene populations than in the older ones (Fig. 7e). The Triassic zircons tend to have lower Th/U than either the Permian or the Jurassic populations. The lowest U and Th contents are found in brecciated and igneous/metamorphic zircons. We interpret this as one effect of the metamorphism and reworking of these complex zircons.

The analyzed zircons have been classified in terms of their parental magmas using the trace-element criteria of Belousova et al. (2001). Nearly all classify as derived from either carbonatite (51/90) or low-Si granoids (36/90); three grains classify as kimberlitic zircons. This pattern is typical, in our experience, of mantle-derived zircons in continental settings, and probably reflects crystallization of zircon from late differentiates of mafic magmas. In the plot of Y vs U/Yb suggested by Grimes et al. (2007, 2015) most of the grains fall into the field of “continental zircon” as distinct from “ocean-crust zircon”, although a significant proportion fall outside the suggested fields due to their low U at low Y contents (Fig. 7f). These anomalous grains largely comprise the brecciated and igneous/metageneric MCZ grains. In the U/Yb-Nb/Yb plot of Grimes et al. (2015) (Fig. 7g) nearly all of the analyzed zircons, regardless of age, fall in the field of Ocean-Island zircons, within the mantle array.

2.3. Hafnium-isotope compositions

Hafnium-isotope data for >350 grains, mainly from the vents, are given in Table A3. The pre-Cretaceous zircons show a range of 176Hf/177Hf from ca 0.2828 to 0.2829 (Fig. 8a), corresponding to εHf values between +5 and +10 for most grains. A plot of 176Hf/177Hf vs age (Fig. 8b) shows a broad trend of increasing 176Hf/177Hf from Permian through Jurassic time, with a mean slope corresponding to 176Lu/177Hf = ca 0.038, essentially parallel to the Depleted Mantle curve. This trend is independent of the morphological classification, and is visible in the individual populations from the vents of Rakefet and Bat Shelomo, for which sufficient data are available. It is typical of mantle-derived zircons from areas such as Siberia and Australia. A similar, but steeper, trend has been recognized in kimberlitic zircons from southern Africa (Fig. 8b; Woodhead et al., 2017).

Most of the Cretaceous zircons have 176Hf/177Hf ratios within the same narrow range observed in the Permian to Jurassic suites (Fig. 9a), and thus have lower εHf values. Many of the Miocene-Pliocene zircons have 176Hf/177Hf within the range covered by the main populations of Cretaceous to Permian zircons (Fig. 9b), but a significant number have significantly more radiogenic Hf, extending toward the Depleted Mantle reference line.

A pattern of a decrease in age without a change in Hf-isotope composition may reflect Pb loss due to later thermal disturbance. In the present case, the situation appears to be more complex. In eight of fourteen core-rim pairs with clearly-defined rims that are significantly younger than the core of the grain, the two zones have identical to nearly identical Hf-isotope compositions (Fig. 9c). Six pairs have rims with significantly higher 176Hf/177Hf than the cores, including one grain with a Jurassic core and a Cretaceous rim; the tie-lines between core and rim in these cases mimic the overall very broad trend to more radiogenic Hf with time (Fig. 9a). There is only one case in which the rim has lower 176Hf/177Hf than the core. The microstructures, combined with the Hf-isotope evidence, suggest that the rims represent growth of new zircon, remobilized from the existing heterogeneous reservoir, rather than simple Pb loss.

2.4. Oxygen-isotope compositions

Oxygen-isotope analyses of 45 spots on 15 zircon grains from the Rakefet volcanic vent (Table A1) cover a range in δ18O from 3.7 to 5.9‰ relative to Vienna Standard Mean Ocean Water (SMOW), with a mean of 5.0 ± 0.45‰ (1 s). One grain gives a very high δ18O = −12.8‰, which is not considered further. Most analyses are within the accepted “mantle range” of 5.6 ± 0.6‰ (2 s), but the lowest values lie outside of this range. Intra-grain variation typically is within the 1 s analytical errors (ca 0.3‰), but one brecciated grain shows a range from 4.7 to 5.4‰ (n = 6 spots), and two igneous/metamorphic grains show core-rim differences of 0.4‰ (n = 2) and 0.9‰ (n = 2), with the rim lighter in each case. The isotopically heaviest oxygen is found in the grains with clearly igneous structures (mean δ18O = 5.5‰; n = 3) and basal spots in igneous/metamorphic grains. The brecciated grains have a
Fig. 6. Extended trace-element patterns of selected zircons by location, and colour-coded by age. Migdal-Ha-Emeq and Kishon River represent detrital (alluvial) zircons; all others are sampled from vents and associated pyroclastics as indicated.
Fig. 7. Plots of trace-element parameters vs age of Mount Carmel zircons. a). La/Sm vs age; b). Eu anomaly vs age; c). Ce anomaly vs age; d). $fO_2$ vs age; e). Th/U vs age; f). discriminant plot relating zircon chemistry to tectonic setting (Grimes et al., 2007); g). discriminant plot relating zircon chemistry to tectonic setting (Grimes et al., 2015). OI, ocean island volcanics. Sloping box encloses zircons from mantle-derived rocks.
mean $\delta^{18}O = 4.8‰$ ($n = 7$, range 3.7–5.1‰); similar low values are found in the igneous/metamorphic population, where averages for individual grains range from 5.3 to 3.8‰. These patterns suggest that the process of metamorphic homogenization and brecciation of primary igneous grains involved fluids with oxygen that was isotopically somewhat lighter than the "mantle average".

3. Discussion

3.1. Younger basalts

The zircon data provide information on the ages and sources of the Lower Basalts and the Cover Basalts in the drainage of the Kishon river, mainly the Yizre‘el Valley. The youngest population of zircons, assumed to represent the Cover Basalts, has $206\text{Pb}/238\text{U}$ ages ranging from $3.6 \pm 0.2$ to $5.5 \pm 0.4$ Ma (mean $4.2 \pm 0.6$ Ma, $1\sigma$, $n = 11$), corresponding to the Lower Pliocene (Fig. 5d). Another major peak at 9–13.5 Ma (mean $11.3 \pm 1.2$ Ma, $n = 22$; Mid- to Upper Miocene) is interpreted as the age of the Lower Basalts in the Yizre‘el Valley; ages back to 17.5 Ma are recognized in the Lower Galilee across the watershed at the head of the Yizre‘el Valley. These ages correspond well with the magmatic episodes defined by Shaliv (1991), Weinstein (2000) and Rozenbaum et al. (2016). The apparent scarcity of magmatic activity between about 9 and 5 Ma in this area also was previously recognized by Shaliv (1991). Zircons in each of these age groups show a large spread in $\varepsilon^{176}\text{Hf}$, ranging from low values similar to the Cretaceous-Permian zircons, up to values closer to the Depleted Mantle (Fig. 9a,b). This range suggests that the late Miocene to Pliocene magmas were derived from the asthenosphere, but also interacted with the SCLM and older underplated basalts (see below) during their ascent.

3.2. Ages of cretaceous volcanic centres

Segev and Rybakov (2011) recognized 5 phases of Cretaceous magmatic activity, interpreted as reflecting the impingement of a mantle plume centered beneath northern Israel, and that caused a major updoming across Lebanon, Syria and the Sinai. The “Tayasir Volcanics” are dated between 141 and 134 Ma (Lower Cretaceous), and basalts of this age have been reported from deep drillings around Mt. Carmel. This age range is poorly represented in our dataset (Fig. 5b), though three grains are similar (133 ± 2 Ma) or slightly older (147–146 Ma). The next episode occurs in lower Aptian time (125–123 Ma; Elije Volcanics); it also is reported from drill cores beneath Mt. Carmel; but is absent in our dataset. The third episode, the late Aptian-Albian Ramon...
volcanics, is represented by a group of 7 zircons with ages ranging from 116 ± 6 Ma to 112 ± 10 Ma. The “Carmel Volcanics”, defined as occurring between ca 100–95 Ma (Table 1), are well-represented in our dataset. The younger Campanian (ca 83–72 Ma) volcanism is represented by a scatter of zircons from the alluvial samples.

The ages of the youngest zircons from each volcanic center can be compared with available K–Ar and Ar–Ar ages (Segev, 2000, 2002, 2009; Segev et al., 2005; Segev and Sass, 2009) (Table 1) to provide further constraints on the age of volcanism and the development of the magmatic system. The Kerem Maharal magmatism, which represents the earliest known eruptions in the Mount Carmel area, has been dated to 99.0 ± 1.0 Ma using amphibole megacrysts (Segev, 2009). Two of the zircon xenocrysts analyzed here give slightly older ages (107 ± 3, 104 ± 3 Ma), while one is significantly younger (90 ± 2 Ma). The Rakefet basalt, which overflies the tuffs sampled for this zircon study (Table 1) has been assigned an average age of 96.7 ± 2.5 Ma (Segev, 2009). Whole-rock Ar–Ar ages on a nearby (E. Muhrqa) basalt are similar (96.2 ± 2.5 Ma; Segev, 2002). Our large dataset (n = 100) from Rakefet contains only one Cretaceous zircon, dated at 93 ± 3 Ma. Two Ar–Ar analyses on an amphibole megacryst from Tavasim (Segev, 2009) give ages of 99.3 ± 0.6 and 98.0 ± 0.5 Ma; both are only slightly younger than the ages of two zircons (102 ± 2 and 101 ± 1 Ma), but as at Kerem Maharal one zircon (84 ± 8 Ma) is younger than the accepted eruption age. The Bat Shelomo volcano has been dated by Ar–Ar analysis of amphibole to be 81.6 ± 0.8 Ma (Segev, 2002); our dataset contains one Cretaceous zircon (99 ± 4 Ma) but two Miocene zircons (13.1 ± 0.3, 12.0 ± 0.3 Ma).

In several of these cases, the youngest zircons are slightly older than the accepted K–Ar and Ar–Ar data. This is consistent with the idea that the zircons are xenocrysts, rather than phenocrysts, in their host magma, and thus must be older than the erupted basalt. However, we see several instances of zircons that are significantly younger than the accepted eruption ages. These may suggest the existence of unrecognized eruption episodes, either in the same area as the sampling, or even re-using the same volcanic plumbing system as the earlier, main eruptions. In some cases, this might be enough to affect K–Ar or Ar–Ar ages of amphiboles; this might explain, for example, the anomalously young age for Bat Shelomo. However, the current datasets from all of the vents except Rakefet are small, and further work on samples from the other vents is needed to clarify these questions.

3.3. Source(s) of the zircon xenocrysts

In the samples from the vents, a large proportion of the zircons have ages corresponding to the Gavim volcanics (Middle Permian to Upper Triassic, 265–220 Ma), and another large population corresponding to the Asher volcanics (Upper Triassic to Lower Jurassic, 206–188 Ma). The Hf and O isotopes of these zircons indicate that their parental magmas were mantle-derived. As noted above, it is most likely that such zircons have crystallized from late differentiates of mantle-derived mafic magmas. Their REE patterns show little obvious evidence of a garnet signature, while the presence of negative Eu anomalies in the most magmatic-looking zircons suggests co-crystallization with plagioclase. This limits the depth of crystallization to <30 km; we suggest that the most likely location for these underplated basalts, where they could cool slowly to produce large grains of cpx, amphibole and zircon, would be at the density filter represented by the crust-mantle boundary (25–20 km in this area; Segev and Rybakov, 2011).

The microstructures of the Permian to Jurassic zircons appear to reflect a complex history of igneous crystallization, metamorphic recrystallization, resetting, and overgrowth. There is a general increase in the proportion of reworked zircons with time (Fig. 9a), but clearly igneous grains are present throughout the Permian-Jurassic population. The Hf-isotope data scatter considerably, but overall define a broad trend of increasing 179Hf/177Hf, with decreasing age. The slope of this trend is essentially parallel to that of the Depleted Mantle reference curve;
of these patterns represent zircons that have been disturbed during re-crystallization, leading to high and irregular REE patterns, and especially high LREE/MREE.

The mid-Triassic melts (230–210 Ma) are similar to most of the older ones, but show less scatter in the REE patterns and smaller Ti anomalies. [Th] and [U] are generally lower than in the older melts, but the range in these contents is similar. The Triassic-Jurassic (210–175 Ma) melts, on the other hand, show considerable diversity relative to the older ones. Some show depletion in HREE, leading to humped patterns; others show either greater or lesser depletion in LREE. [Th], [U], [Nb] and [Ta] are higher overall, and the range in Th/U is greater; Ti anomalies, both positive and negative, also show a greater range.

The flat MREE-HREE patterns with LREE depletion are reminiscent of the REE patterns seen in e.g. N-MORB, but the anomalies in Ce and Ti, and the high contents of Nb, Th and U are not typical of MORB. It is, in fact, difficult to find primary mantle melts with similar patterns. However, such patterns might develop through extended fractional crystallization of magmas with less pronounced REE patterns. The basalts of N. Israel and adjacent areas, from Jurassic through lower Cretaceous and Neogene time, are alkali basalts with remarkably constant trace-element patterns and Sr—Nd isotopes (Stein and Hofmann, 1992; Weinstein et al., 2006). Their trace-element patterns (Fig. 10a) have mildly negative slopes from Yb to La, with [Th], [U], [Nb] and [Ta] at ca 100 times chondrites. They have no anomalies in Eu, Ce or Ti; CN-normalized P/Yb is ca 0.5.

Compared to these probable primary basalts, the strong depletion of LREE in the calculated melts may reflect crystallization of late phases such as apatite and monazite, and the HREE depletion could reflect crystallization of both zircon and xenotime. Removal of apatite also can produce positive Eu anomalies, counteracting the influence of plagioclase crystallization. The depletion of P relative to the HREE (P/Yb ca 0.1) seen in all of the melts, and the wide variation in Th and U, also may be tied to the crystallization of monazite and xenotime, as well asapatite. These comparisons suggest that the calculated melts, and the zircons crystallized from them, represent late-stage differentiates of basaltic melts cooling slowly in deep-seated magma chambers.

The zircons erupted from Upper Cretaceous to Neogene time are broadly similar to the Permo-Jurassic ones in having strong LREE depletion, and high Th—Ta (Fig. 10). However, they show generally more enhanced Ce anomalies, smaller Eu anomalies (Fig. 7b,c), and progressively develop mild HREE depletion relative to the MREE, producing concave-downward REE patterns. There is also a tendency toward deeper negative Ti anomalies, especially in the youngest zircons, and toward Th/U values <1.

3.5. Sampling patterns in the cretaceous volcanism

The group of zircons with ages in the Upper Cretaceous (116–95 Ma) may give an indication of the duration of this episode at depth, even though the range of eruption ages is shorter. As noted above, it is likely that none of the Cretaceous zircons truly date their respective eruption, but rather represent xenocrysts derived from roughly coeval, more differentiated, magmatism at depth. The range of 176Hf/177Hf values in the Cretaceous zircons corresponds closely to that of the Permian-Jurassic populations (Fig. 8), suggesting the Cretaceous magmas were derived from the same reservoir.

The distribution pattern of the older populations in the different Cretaceous vents gives a picture of this reservoir (Fig. 11). The Permian-Triassic (280–220 Ma) zircons are most abundant in the main vents (Rakefet, Bat Shelomo, Makura and the Ein Hashofet centre in the Ramot Menashe syncline). The Triassic-Jurassic (Asher) zircons are found mixed with Permian-Triassic ones in Ein Hashofet and Rakefet, but make up the bulk of the sample in the northern bodies (Beit Oren, Karem Maharal and Har Alon); the younger zircons are found mainly in Karem Maharal, Tavassim and Migdal Ha-Emeq on the western side of Mt. Carmel. This pattern suggests that the Permo-Triassic magmatism built up a “pillow” beneath the core of the Mt. Carmel-Menashe area, so that the magmas of the younger episodes tended to intrude toward the margins of this central area (Fig. 12).

Segev and Rybakov (2011) used data from deep drillings and geophysical surveys (primarily magnetics and gravity) to outline areas showing “significant influence of magmatic bodies”. In this synthesis, the Late Triassic-Jurassic magmatism is shown extending along the current coastline both NE and SW of the Mt. Carmel area, while Permo-Triassic magmatism is isolated farther south. However, the data presented in this paper indicate that the Permo-Triassic magmatism also was significant beneath the Mt. Carmel area and probably contributes to the complex geophysical signature of the area.

The Permian-Jurassic volcanic rocks reflect the formation of the Levant continental margin by rifting during the breakup of Gondwana, which produced the Neotethys and Mesotethys oceans. Segev (2000) listed 17 globally synchronous short-term volcanic cycles (200 to 5
Ma) that are expressed in this wider region, and proposed that they represent pulses of plume activity. An alternative interpretation would be that they reflect periods of tectonic re-adjustment during which thinning and mantle upwelling led to the production of new melts from the proposed fossil plume head of Stein and Hofmann (1992), while coeval rifting along old fracture systems allowed the magmas access to the upper lithosphere and the surface.

In this framework, the concentration of the Cenomanian volcanism (97–80 Ma) into the Mount Carmel area may reflect its extensive network of translithospheric and crustal-scale high-angle faults, most of which were reactivated in Miocene time. We cannot know the real extent of the Permian-Jurassic underplating beneath this area; it may extend far beyond the 150–300 km$^2$ sampled by the known Cretaceous vents, and connect with the area identified by Segev and Rybakov (2011). However, it seems likely that the same set of translithospheric structures, dating back to the initial rifting from Gondwana, has localized magmatism beneath this area for at least 250–300 million years.

4. Conclusions

Zircon xenocrysts from Cretaceous pyroclastic deposits on Mt. Carmel and associated alluvial deposits document extensive magmatic activity beneath the area during Permo-Triassic and Late Triassic-Jurassic time. The morphology and internal structures of the grains suggest a magmatic origin, followed by recrystallization, overgrowth and brecciation. Trace-element data indicate that the parental magmas were alkaline to carbonatitic, and probably represent residual differentiates of basaltic melts. Hf-isotope ratios indicate derivation of these melts from an enriched mantle (OIB-like) source, probably mid- to Neoproterozoic in age. This may represent the plume head postulated by Stein and co-workers as the source of basalts across the region. The distribution of the different age populations across Mt. Carmel and neighbouring areas suggests that the earliest intrusions built up an underplated “pillow”, probably near the crust-mantle boundary, and that the emplacement of the later intrusions occurred primarily around the edges of the pillow.
The Permain to Jurassic zircons record the repeated intrusion of magmas derived from a common source, and the (repeated?) recrystallization of older zircons in the presence of fluids with relatively light O-isotopes. The Cretaceous zircons have HF-isotope ratios within the range of the older zircons in the presence of fluids with relatively light O-isotopes. The derived zircons suggest the derivation of magmas both from the older “plume-related” source and from a more primitive convecting mantle.

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Appendix A. Supplementary data

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References