Thermotectonic history of SE China since the Late Mesozoic: insights from detailed thermochronological studies of Hong Kong

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Abstract: The late Mesozoic Yanshanian volcanic arc affected an extensive region of SE China, but the conclusion of magmatism and later evolution are not fully understood. Widespread Yanshanian ignimbrites and their contemporaneous granites exposed in Hong Kong represent a microcosm of this magmatic arc. To constrain the post-magmatic thermal history of the region, we present zircon and apatite fission-track analyses from these rocks. Double dating using laser ablation inductively coupled plasma mass spectrometry U–Pb and fission-track techniques on detrital zircons from post-volcanic Cretaceous sediments is used to further constrain the tectonothermal evolution. The resulting dataset and thermal modelling suggest that the igneous rocks and Cretaceous sediments together experienced post-emplacement or post-depositional heating to >250 °C, subsequently cooling through 120–60 °C after c. 80 Ma. The heating reflects the combined effects of an enhanced geothermal gradient and burial. We interpret the enhanced gradient to represent continuing Yanshanian magmatic activity until c. 100–80 Ma, much later than previously considered. Our data also indicate a long-term, slow cooling (c. 1 °C myr⁻¹) since the early Cenozoic, linked to c. 2–3 km of erosion-driven exhumation. The thermotectonic history of Hong Kong reflects the mid-Cretaceous transition of SE China from an active to a passive margin bordered by marginal basins that formed in the early Cenozoic.

Supplementary material: Descriptions of samples, operating conditions of the laser ablation inductively coupled plasma mass spectrometry system and the full dataset of U–Pb dating of detrital zircons are available at www.geolsoc.org.uk/SUP18750.

During the late Mesozoic era, an active magmatic arc was present in SE China during the Yanshanian Orogeny. Widespread and voluminous magmatism and volcanism occurred along the southern margin of the Eurasian landmass, now represented by the 1300 km long and 400 km wide SE China Magmatic Belt (Fig. 1; Zhou et al. 2006; Li & Li 2007). SE China then underwent a transition from an active continental margin to a passive tectonic setting through multiple phases of crustal extension after the late Mesozoic (Gilder et al. 1991; Chan et al. 2010). Numerous studies have investigated this evolution through geochemical and isotopic analyses on post-Yanshanian magmatic rocks (e.g. Chung et al. 1997; Huang et al. 2013), facies and provenance analysis on offshore sediments (e.g. Wu 1994; Li et al. 2012), detailed structural mapping (e.g. Chan et al. 2010), analysis of onshore and offshore rift basins (e.g. Gilder et al. 1991; Zhou et al. 1995; Shu et al. 2009), and modelling of geophysical data (e.g. Hayes et al. 1995; Nissen et al. 1995a,b; Shi & Li 2012). Despite this level of detail, the timing of shutdown of silicic magmatism, collapse of the magmatic arc and evolution of marginal basins is not well constrained.

Studies employing low-temperature thermochronology have been carried out in Guangdong Province, the northern neighbour of Hong Kong, to understand the thermal evolution of the SE China Magmatic Belt. Li et al. (2005) reported zircon fission-track (ZFT) and apatite fission-track (AFT) ages (c. 97–133 Ma and c. 43–68 Ma, respectively) from western Guangdong (Fig. 1), and proposed that the region had experienced exhumation of >4 km since c. 130 Ma. Yan et al. (2009) used ZFT, AFT and (U–Th)/He techniques to reconstruct the exhumation history of granitic rocks along a NW–SE transect in central Guangdong Province (Fig. 1). They proposed a three-stage cooling history of the granites based on inverse modeling of the data: a rapid initial cooling during the Late Cretaceous, followed by a thermally stable period at around 60 °C from 60 to 15 Ma, and a final stage of cooling from c. 15–10 Ma to the present day. A subsequent ZFT and AFT study undertaken by Tsang (2010) along the coastal region of Guangdong Province (Fig. 1), however, implied the presence of cooling periods that were inconsistent with those determined by Yan et al. (2009). Tsang (2010) reported ZFT ages (c. 165–77 Ma) and AFT ages (c. 46–0.8 Ma) from a wide range of lithologies, and postulated that there were at least two episodes of cooling during the Cenozoic: a rapid cooling at c. 30 Ma and a slower cooling at c. 15–10 Ma. The results of these studies were, in general, inconsistent with each other.

Hong Kong, located on the coastal margin of SE China, represents a microcosm of the SE China Magmatic Belt. The geology, well understood through detailed mapping, is dominated by late Mesozoic volcanic rocks and sub-volcanic plutons, now exposed as a result of crustal uplift and erosion (Sewell et al. 2012a). The post-magmatic evolution has generally been considered to be
fairly simple, with the development of a series of Cretaceous fault-bounded basins followed by a long tectonically stable period (Lai 1985; Sewell et al. 2000). However, some geochronological evidence points to possible periods of later tectonic and/or thermal activity. Campbell & Sewell (2005) reported ⁴⁰Ar–³⁹Ar (biotite and whole-rock) ages of mafic dykes ranging between 106 and 75 Ma, noting that the ages were much younger than the U–Pb zircon ages of apparently coeval magmatic rocks. They suggested that the ⁴⁰Ar–³⁹Ar system might have been reset and that the ages probably represent a thermal event around 100–90 Ma in the region. Additionally, Campbell & Sewell (2005) also identified possible faulting events at 34, 10 and 3–4 Ma by dating sheared lithologies on major faults using ⁴⁰Ar–³⁹Ar methods on whole-rock and minerals (feldspar and pyrite).

To investigate the post-volcanic thermal–tectonic history hinted at by the ⁴⁰Ar–³⁹Ar ages, we have employed low-temperature thermochronology on the Middle Jurassic to Early Cretaceous volcanic–plutonic assemblages of Hong Kong. The only fission-track (FT) studies previously reported (Nau & Yim 1988) were based on cannic–plutonic assemblages of Hong Kong. The only fission-track chronology on the Middle Jurassic to Early Cretaceous volcanic rocks of Hong Kong. Isotopic and geochemical studies show that the basement rocks of the South China Block consist mainly of Archaean, and Palaeo- to Mesoproterozoic rocks (Jahn et al. 1990; Li et al. 1994; Lai 1993; Lee & Lawver 1994). Eruption of Palaeocene–Eocene basalts probably signalled the initiation of rifting in the coastal region (e.g. Chung et al. 2008–2013), prior to the beginning of sea-floor spreading in the South China Sea.

Geological setting

Regional geology

SE China comprises two major crustal blocks: the Yangtze Block in the north and the Cathaysia Block in the south (Fig. 1; Hsü et al. 1990; Jahn et al. 1990), collectively known as the South China Block. Isotopic and geochemical studies show that the basement rocks of the South China Block consist mainly of Archaean, and Palaeo- to Mesoproterozoic rocks (Jahn et al. 1990; Li et al. 1994; Lai 1993; Lee & Lawver 1994). Overlying this basement are Palaeozoic sedimentary sequences, an extensive Mesozoic igneous province (SE China Magmatic Belt, Fig. 1) forming the southeastern coastal belt, and relatively limited Mesozoic and Cenozoic sediments (Bureau of Geology & Mineral Resources of Guangdong Province 1988). However, this simplified picture is rendered very complex owing to multiple phases of extension and rifting since the late Mesozoic.

Two Mesozoic orogenies, the Triassic Indosinan and the Middle Jurassic to Late Cretaceous Yanshanian, have been identified in SE China. Over 90% of the exposed igneous rocks were formed during the prolonged Yanshanian magmatism (Zhou et al. 2006), which lasted from the Middle Jurassic to the mid- to Late Cretaceous (c. 180–80 Ma: Li 2000; Zhou & Li 2000; Li & Li 2007). The Yanshanian magmatism has been generally linked to the subduction of the palaeo-Pacific Plate (Izanagi Plate) beneath the Eurasian Plate (e.g. Jahn 1974; Charvet et al. 1994; Wang et al. 2011), although some researchers (e.g. Li et al. 2004; Chen et al. 2008) argued that the extensive silicic magmatism was instead related to rifting post-dating the Indosinan Orogeny.

Post-Yanshanian continental extension and rifting was considered to be associated with the retreat of the subduction zone in the late Mesozoic (Ren et al. 2002; Shi & Li 2012). The extensional tectonics was manifested in the development of late Mesozoic terrestrial basins, filled with coarse, immature volcanioclastic-rich sediments (Zhou et al. 1995; Chan et al. 2010; Shi & Li 2012). Oil-bearing Eocene basins subsequently developed at the attenuated continental margin in SE China (Chung et al. 1997). The continental extension was followed by opening of the South China Sea during the Oligocene to Miocene (Ru & Pigott 1986; Briais et al. 1993; Lee & Lawver 1994). Eruption of Palaeocene–Eocene basalts probably signalled the initiation of rifting in the coastal region (e.g. Chung et al. 1994; Yan et al. 2006; Wang et al. 2012; Huang et al. 2013), prior to the beginning of sea-floor spreading in the South China Sea.

Geology of Hong Kong

About 85% of the land area is made up of late Mesozoic Yanshanian magmatic rocks (Fig. 2), including the volcanic and intrusive products of four temporally distinct magmatic episodes at 164–160, 148–146, 143 and 141–140 Ma (table 3 of Sewell et al. 2012b). The volcanic rocks include dacite to rhyolitic ignimbrites and lavas with intercalated tuffaceous sedimentary units, and the intrusive rocks are granodiorite, granite, quartz monzonite and minor dykes,

we are able to establish the thermotectonic history in light of the local geological framework.

The specific objectives of this study were therefore to determine the post-magmatic thermal histories by (1) constraining temporally the burial and exhumation of the Mesozoic volcanic complexes in Hong Kong, (2) assessing whether there was significant sedimentation across the region after magmatism had ended and (3) identifying the timing of regional inversion, rock uplift and exhumation.

Geological setting

Regional geology
Fig. 2. Geology of Hong Kong (modified after Sewell et al. 2000) with sample locations and ages. Isotope dilution thermal ionization mass spectrometry (ID-TIMS) U–Pb zircon ages after Davis et al. (1997), Campbell et al. (2007) and Sewell et al. (2012b). All U–Pb zircon ages, and ZFT and AFT ages are in Ma with a 2σ error.
of various compositions (Sewell & Campbell 1997; Sewell et al. 2000, 2012a). The volcanic and plutonic rocks are considered to be largely eogenetic and to represent ancient caldera-type large-scale silicic magmatic systems (Campbell & Sewell 1997). Sewell et al. (2012a) have recently proposed a tilted nested caldera model for the two Cretaceous magmatic assemblages (143 and 141–140 Ma) in SE Hong Kong.

Four post-volcanic sedimentary units have been identified: the Cretaceous Pat Sin Leng, Port Island and Kat O formations (collectively known as the Miers Bay Group; Fig. 2), and the Eocene Ping Chau Formation (Fig. 2; Sewell et al. 2000). The Cretaceous units are dominated by non-fossiliferous, reddish sandstone, siltstone and conglomerate deposited in fluvial environments (Jones 1995, 1996). All units unconformably overlie the Middle Jurassic to Lower Cretaceous volcanic rocks in NE Hong Kong: their stratigraphic ages are only loosely constrained by correlation with the underlying volcanic sequences. The Eocene Ping Chau Formation consists of dolomitic and calcareous lacustrine siltstone, bearing a diverse assemblage of fossil plants and insects (Lai et al. 1996). This formation represents the youngest preserved rock formation in Hong Kong, and a major hiatus exists between these indurated Eocene rocks and unconsolidated Quaternary sediments (Sewell et al. 2000).

**Methods**

**Fission-track analysis**

Representative samples of both volcanic and plutonic units from all four late Mesozoic magmatic episodes were collected from a wide geographical distribution, especially across major fault structures in Hong Kong. In addition, rock samples from all four post-volcanic sedimentary units were collected from their type localities. Sample locations are shown in Figure 2.

Bulk-rock samples were processed using mechanical crushing and conventional mineral separation techniques (through heavy liquids and magnetic separation). The sedimentary samples generally yielded insufficient detrital apatite grains for fission-track analysis. Aliquots of apatite and zircon were mounted in epoxy resin and Teflon sheet, respectively. For the detrital zircons, two mounts per sample were made to allow for etching at different lengths of time to adjust for the possibility of a range of ages. Mineral grains were then polished to expose the internal surfaces. The apatite grains were etched with 5M HNO₃ at 20 °C for 20 s and the zircons with eutectic KOH–NaOH melt at c. 210 °C for 24–50 h. The external detector method was used throughout (Gleadow 1981; Hurford & Green 1983). The neutron irradiation of the samples was carried out at the Oregon State University Radiation Facility, USA (samples analysed by D.L.K.T.) and in the FRM 11 thermal neutron facility at the University of Munich in Germany (samples analysed by A.C.). Dosimeter glasses CN1 (D.L.K.T.) and CN2 (A.C.) for zircon and CN5 for apatite were used. The spontaneous and induced fission tracks of single grains were counted under a Zeiss Axiosoplan microscope with a computer-controlled stage and a total magnification of 1250× (A.C.), or with a Zeiss MM+ microscope at a total magnification of 1000× (D.L.K.T.). The zeta (ζ) calibration method and IUGS recommended age standards (Hurford & Green 1983; Hurford 1990) were used throughout. Zircon CN1-ζ is 129±13 (D.L.K.T.) and CN2-ζ is 127±4 (A.C.) and apatite CN5-ζ is 339±5 (A.C.). The reported ages are central ages (with 2σ error) calculated using the TRACKKEY program (Dunkl 2002). The lengths of horizontal confined fission tracks were measured by A.C. Using FASTTRACKS software (Gleadow et al. 2012) the lengths of confined inclined tracks were additionally determined (D.L.K.T.). To infer the time-temperature history of the late Mesozoic igneous rocks, inverse modelling was carried out using the AFT ages and the confined track-length measurements, according to the procedures of Ketcham (2005, HeFTy program version 1.8.0).

**U–Pb dating of detrital zircon using LA-ICP-MS**

Five zircon suites from the post-volcanic sedimentary units that had already been dated by the fission-track method were analysed for U–Pb ages using LA-ICP-MS techniques at the School of Earth Sciences, University of Melbourne using an Agilent 7700× coupled to a Helex ArF 193 nm laser ablation system. Ablation was conducted in a He atmosphere, and the sample and He were mixed with Ar en route to the mass spectrometer (see Woodhead et al. 2007). Baseline-corrected 238U and 206Pb counts for reference zircon 91500 (Wiedenbeck et al. 2004) were typically c. 100000 and 17000 counts per second, respectively. All data were reduced using the U–Pb-Geochronology3 data reduction scheme in iolite (Paton et al. 2010, 2011). Zircon 91500 was used as a primary standard to correct for down-hole fractionation and U–Pb ratio normalization. Temora 2 (417 Ma, Black et al. 2004) and Plesovice (337 Ma, Slama et al. 2008) reference zircons were also measured concurrently as secondary standards and produced ages in the range of 423±27 Ma (2σ, n=7) and 347±10 Ma (2σ, n=10), respectively, with no outlier rejection. U–Pb age components of the detrital zircons were determined using DensityPlotter, which allows visualization of both the kernel density estimate and probability density plot (Vermeech 2012).

**Results**

**Apatite and zircon fission-track analysis of magmatic rocks**

The ZFT and AFT data are summarized in Tables 1 and 2, respectively. The ZFT ages for granites range between c. 109 and 60 Ma and for volcanic rocks between c. 140 and 69 Ma (Fig. 3a). Regardless of geographical location, all of the volcanic rocks and granites yielded ZFT ages younger (by 20 to >100 myr) than their eruption or emplacement ages (Fig. 3a). The AFT ages for all samples range from 83 to 40 Ma, except for two outliers of 28 and 23 Ma (Fig. 3b). They are on average 30 myr younger than their eruption or emplacement ages (Fig. 3a). The AFT ages for all samples range from 83 to 40 Ma, except for two outliers of 28 and 23 Ma (Fig. 3b). They are on average 30 myr younger than their equivalent ZFT ages and also show no apparent pattern with respect to the geographical distribution. The sample elevations range from 0 to 700 m above sea level. No noticeable correlation is observed between the FT ages and the elevations of either volcanic or granitic samples from different regions of Hong Kong (Fig. 3c and d).

**Detrital U–Pb zircon dating using LA-ICP-MS**

The U–Pb age data are of good quality for provenance identification purpose, even though some of the detrital grains contain low uranium concentrations. The Pat Sin Leng Formation (sample HK13344) yields euhedral detrital zircons that contain age components of c. 120, c. 142 and c. 162 Ma, with another possible peak
| Sample ID | Lithology       | U–Pb age (Ma) | Coordinates (Easting, Northing) | Altitude (m) | Number of grains | $\rho_d$ ($N_d$ counted) | $\rho_s$ ($N_s$ counted) | $\rho_i$ ($N_i$ counted) | $U$ (ppm) | $P(\chi^2)$ % (dispersion) | Central age ± 2σ (Ma) |
|-----------|-----------------|---------------|--------------------------------|--------------|------------------|------------------------|------------------------|------------------------|----------|----------------------------|----------------------|
| HK8758    | Quartz monzonite| 140.4 ± 0.3   | 22°12′47.62″, 113°55′06.16″    | 65           | 25               | 5.14 (3002)           | 112.6 (1554)           | 48.8 (674)            | 315      | 25.0 (0.12)                  | 76.1 ± 17.6          |
| HK8353    | Granite         | <143.7 ± 0.3  | 22°13′04.11″, 113°59′58.20″    | 2            | 15               | 5.18 (2874)           | 108.1 (1918)           | 32.4 (575)            | 184      | 9.0 (0.12)                   | 109.1 ± 14.6         |
| HK8754    | Granite         | 159.3 ± 0.3   | 22°25′00.79″, 113°59′57.64″    | 108          | 8                | 5.18 (2874)           | 119.1 (941)            | 59.0 (466)            | 340      | 10.2 (0.11)                  | 65.7 ± 10.2          |
| HK11042   | Granite         | 140.4 ± 0.2   | 22°18′37.85″, 114°10′33.32″    | 35           | 20               | 5.10 (3002)           | 143.6 (2412)           | 68.3 (1147)           | 452      | 22.9 (0.09)                  | 68.1 ± 12.4          |
| HK11831   | Feldspar porphyry| 146.5 ± 0.2  | 22°17′36.41″, 114°00′28.95″    | 140          | 15               | 5.48 (3794)           | 108.7 (2335)           | 36.8 (790)            | 196      | 0.0 (0.24)                   | 102.0 ± 16.8         |
| HK11839   | Granite         | 146.2 ± 0.2   | 22°22′20.22″, 114°10′29.91″    | 80           | 14               | 5.18 (2874)           | 104.1 (1760)           | 35.2 (595)            | 199      | 29.8 (0.06)                  | 96.5 ± 12.0          |
| HK12023   | Granite         | 140.6 ± 0.3   | 22°12′58.10″, 114°08′02.49″    | 5            | 17               | 5.20 (3002)           | 158.5 (1585)           | 89.1 (891)            | 544      | 7.5 (0.10)                   | 60.4 ± 13.8          |
| HK13278   | Granite         | 146.4 ± 0.1   | 22°20′52.57″, 114°10′08.23″    | 415          | 11               | 5.48 (3794)           | 152.1 (1138)           | 55.9 (418)            | 290      | 1.1 (0.15)                   | 94.0 ± 12.6          |
| HK13277   | Trachydacite   | 141.2 ± 0.1   | 22°17′08.17″, 114°17′24.67″    | 5            | 25               | 5.03 (3002)           | 98.8 (1869)            | 0.262 (678)           | 231      | 7.9 (0.13)                   | 93.6 ± 21.2          |
| HK12025   | Ash tuff       | 164.2 ± 0.3   | 22°16′10.29″, 113°53′08.25″    | 2            | 12               | 5.18 (2874)           | 114.0 (1880)           | 36.6 (603)            | 213      | 20.5 (0.09)                  | 101.2 ± 15.6         |
| HK12026   | Rhyodacite     | 164.1 ± 0.3   | 22°31′31.77″, 114°17′10.14″    | 2            | 15               | 5.48 (3794)           | 117.3 (2746)           | 28.8 (679)            | 157      | 83.5 (0.00)                  | 140.0 ± 12.8         |
| HK13275   | Crystal tuff   | 143.0 ± 0.2   | 22°24′23.33″, 114°07′19.04″    | 130          | 15               | 5.48 (3794)           | 110.3 (2110)           | 38.9 (799)            | 209      | 0.12 (0.18)                  | 101.9 ± 13.0         |
| HK13339   | Crystal tuff   | No U–Pb age;  | 22°29′19.49″, 114°09′10.16″    | 112          | 15               | 5.48 (3794)           | 183.0 (3293)           | 47.7 (1274)           | 253      | 0.0 (0.22)                   | 86.9 ± 13.0          |

$\rho_s$ and $\rho_i$ represent sample spontaneous and induced track densities; $\rho_d$ is the standard track density; $\rho_p$, $\rho_s$, and $\rho_i$ are in units of $\times 10^5$ tracks cm$^{-2}$. Ages were calculated using dosimeter glasses CN1 and CN2, where $\zeta_{CN1} (D.L.K.T.) = 129 ± 13$ and $\zeta_{CN2} (A.C.) = 127 ± 4$ (Hurford & Green 1983). $P(\chi^2)$ is probability for obtaining $\chi^2$ value for $n$ degrees of freedom, where $n = $ number of crystals − 1. Central age is a modal age, weighted for different precisions of single crystals (Galbraith 1990). U–Pb ages are based on Davis et al. (1997), Campbell et al. (2007) and Sewell et al. (2012a).
Table 2. Apatite fission-track ages for magmatic rock samples

| Sample ID | Lithology       | U–Pb age (Ma) | Coordinates (Easting, Northing) | Altitude (m) | Number of grains | \( \rho_d \) (Nd counted) | \( \rho_i \) (Nd counted) | \( \rho_s \) (Ns counted) | U (ppm) | \( P(\chi^2) \) % (Dispersion) | Central age ± 2\( \sigma \) (Ma) | Mean track length (µm) (number of tracks measured) | \( \sigma \) (µm) |
|-----------|----------------|---------------|---------------------------------|--------------|------------------|--------------------------|--------------------------|--------------------------|---------|-----------------------------|-----------------------------|---------------------------------|---------|
| **Intrusive rocks** | | | | | | | | | | | | | |
| HK8754    | Granite        | 159.3 ± 0.3  | 22°25′00.79″, 113°59′57.64″ | 108          | 22               | 14.62 (6079)     | 2.7 (140)               | 16.2 (829)               | 15      | 71.7 (0.07) 41.7 ± 7.8       | 13.48 ± 0.7 (4) 1.41           | –                                |         |
| HK8758    | Quartz monzonite | 140.4 ± 0.3  | 22°12′47.62″, 113°55′06.16″ | 65           | 22               | 14.62 (6079)     | 1.9 (151)               | 10.4 (1351)              | 9       | 51.9 (0.07) 46.2 ± 8.6       | 12.91 ± 0.18 (98) 1.78          | –                                |         |
| HK1042    | Granite        | 140.4 ± 0.2  | 22°18′37.85″, 114°10′33.32″ | 35           | 32               | 14.62 (6079)     | 1.8 (207)               | 8.0 (912)                | 7       | 97.6 (0.01) 56.0 ± 9.0       | 12.33 ± 0.22 (78) 1.91          | –                                |         |
| HK8759    | Granite        | 146.2 ± 0.2  | 22°22′20.22″, 114°10′02.91″ | 80           | 30               | 14.62 (6079)     | 2.5 (165)               | 11.2 (751)               | 9       | 47.4 (0.14) 54.7 ± 10.2      | –                                | –                                |         |
| HK12003   | Quartz porphyry | 146.3 ± 0.3  | 22°20′38.11″, 114°02′40.83″ | 20           | 28               | 14.62 (6079)     | 1.0 (103)               | 8.4 (877)                | 8       | 84.2 (0.07) 28.0 ± 6.0       | 14.60 ± 0.70 (3) 1.21          | –                                |         |
| HK12022   | Quartz monzonite | 146.6 ± 0.3  | 22°12′43.11″, 114°14′51.00″ | 190          | 31               | 14.62 (6079)     | 2.2 (263)               | 9.1 (1093)               | 7       | 55.9 (0.07) 59.4 ± 8.6       | 13.24 ± 0.23 (55) 1.68          | –                                |         |
| HK13278   | Granite        | 146.4 ± 0.1  | 22°20′52.57″, 114°10′08.23″ | 415          | 23               | 14.73 (4082)     | 1.4 (132)               | 5.0 (491)                | 4       | 70.2 (0.07) 67.0 ± 13.6      | –                                | –                                |         |
| **Volcanic rocks** | | | | | | | | | | | | | |
| HK11821   | Crystal tuff   | 164.5 ± 0.2  | 22°19′55.47″, 114°01′46.01″ | 0            | 32               | 14.62 (6079)     | 2.7 (240)               | 15.0 (1351)              | 14      | 91.4 (0.00) 43.9 ± 6.4       | 13.09 ± 0.43 (16) 1.74          | –                                |         |
| HK11832   | Crystal tuff   | 142.8 ± 0.2  | 22°26′10.10″, 114°20′09.99″ | 10           | 32               | 14.62 (6079)     | 1.7 (187)               | 6.7 (751)                | 5       | 11.3 (0.27) 67.3 ± 13.4      | 12.81 ± 0.85 1.10                | –                                |         |
| HK11835   | Crystal tuff   | 142.7 ± 0.2  | 22°22′47.12″, 114°15′47.23″ | 65           | 32               | 14.62 (6079)     | 1.8 (175)               | 5.3 (517)                | 4       | 74.3 (0.03) 83.3 ± 15.8      | 13.80 ± 0.23 (52) 1.67          | –                                |         |
| HK11836   | Crystal tuff   | 142.5 ± 0.3  | 22°19′24.24″, 114°16′18.21″ | 10           | 28               | 14.62 (6079)     | 0.9 (62)                | 4.9 (331)                | 4       | 98.3 (0.00) 46.4 ± 13.0      | –                                | –                                |         |
| HK11837   | Crystal tuff   | <164.5 ± 0.7 | 22°24′40.62″, 114°07′00.04″ | 680          | 30               | 14.62 (6079)     | 1.6 (157)               | 8.8 (844)                | 8       | 75.1 (0.02) 45.9 ± 8.2       | 14.37 ± 0.29 (15) 1.14          | –                                |         |
| HK12001   | Vitric tuff    | 140.9 ± 0.2  | 22°21′51.34″, 114°22′29.05″ | 25           | 30               | 14.62 (6079)     | 1.9 (253)               | 7.9 (1074)               | 6       | 48.0 (0.02) 58.1 ± 8.4       | 13.06 ± 0.21 (80) 1.88          | –                                |         |
| HK12076   | Crystal tuff   | 142.8 ± 0.2  | 22°16′29.29″, 113°58′14.90″ | 658          | 22               | 14.73 (4082)     | 1.3 (135)               | 5.3 (552)                | 4       | 11.8 (0.29) 60.1 ± 14.4      | –                                | –                                |         |
| HK13330   | Crystal tuff   | No U–Pb age; 22°15′12.94″, 113°51′10.15″ | 2            | 8               | 13.91 (5023)     | 2.8 (62)               | 9.7 (214)                | 9       | 0.6 (0.48) 72.0 ± 32.2       | –                                | –                                |         |
| HK13332   | Crystal tuff   | No U–Pb age; 22°12′11.71″, 113°52′03.88″ | 2            | 7               | 13.91 (5023)     | 0.7 (17)               | 7.1 (177)                | 6       | 87.9 (0.00) 22.6 ± 11.6      | –                                | –                                |         |

\( \rho_s \) and \( \rho_i \) represent sample spontaneous and induced track densities; \( \rho_d \) is the standard track density; \( \rho_s, \rho_i \), and \( \rho_d \) are in units of \( \times 10^5 \text{ tracks cm}^{-2} \). Ages were calculated using dosimeter glass CN5, where \( \zeta_{CN5 \text{ (A.C.)}} = 339 \pm 5 \) (Hurford & Green 1983). \( P(\chi^2) \) is probability for obtaining \( \chi^2 \) value for \( n \) degrees of freedom, where \( n = \text{number of crystals} - 1 \). Central age is a modal age, weighted for different precisions of single crystals (Galbraith 1990). U–Pb ages are based on Davis et al. (1997), Campbell et al. (2007) and Sewell et al. (2012a).
Fig. 3. Emplacement age v. (a) ZFT age and (b) AFT age; elevation of sample v. (c) ZFT age and (d) AFT age. Open symbols represent volcanic rocks and filled symbols intrusive rocks. Error bars (2σ) for zircon U–Pb ages by ID-TIMS techniques are smaller than the size of symbols in the figure.

Fig. 4. Relative age probability plots of detrital U–Pb zircon ages of post-volcanic sedimentary units. Dashed lines represent kernel density estimate; continuous lines represent probability density plot. Peak age locations and proportions were estimated using on mixture modelling algorithm in DensityPlotter (Vermeesch 2012). Age populations between 100 and 300 Ma, which made up over 90% of all age data, are presented in the figure.
at c. 152 Ma (Fig. 4). Two of 53 analyses yield U–Pb ages of >900 Ma. Detrital zircons from the Port Island Formation (sample HK13329) are dominated by the c. 143 Ma population, with two minor populations at c. 156 and c. 205 Ma identified by DensityPlotter (Fig. 4). Only two out of 71 analyses from sample HK13329 are over 300 Ma. Two samples from the Kat O Formation (samples HK13326 and HK13325) have slightly dissimilar detrital age populations of c. 143, c. 152 and c. 164–160 Ma (Fig. 4). Six out of a total of 94 analyses from this formation yield U–Pb ages over 300 Ma.

The Eocene Ping Chau Formation (sample HK13322) contains mixed rounded and euhedral detrital zircon grains. Only 29 grains were analysed owing to a poor yield from the sample and generally small grain sizes. Of these analysed grains, 10 are of Palaeozoic to Palaeoproterozoic age. The remaining grains fall into age populations of c. 152 and c. 164 Ma (Fig. 4).

**D detrital apatite and zircon fission-track analysis of sedimentary rocks**

All ZFT ages from the Cretaceous sedimentary rocks are younger than the detrital U–Pb ages and all analyses pass the χ2 test indicating a high probability that the source rock signature has been fully removed (Table 3). All ages overlap statistically between 75 and 89 Ma at 2σ levels. This observation implies resetting of the system to ≤250 °C after deposition. Detrital apatite from the Kat O Formation yields an AFT age of about 63 Ma, which again falls within the range of AFT ages of the igneous samples. The Eocene Ping Chau Formation (sample HK13323), in contrast, yielded detrital ZFT grain ages ranging between 70 and 110 Ma (i.e. older than the stratigraphic age). These data imply that the detrital zircons have not been reset after deposition, but reflect primarily a record of the thermal history of the source rocks.

Two of the sedimentary samples (HK13326 and HK13329 from the Port Island and Pat Sin Leng formations, respectively) have yielded zircon He ages, with a closure temperature of c. 180 °C (Reiners 2005), of 93.7 ± 5.7 Ma and 93.0 ± 14.0 Ma, respectively (weighted mean ages with 2σ error; B. Kohn, unpubl. data). These results are statistically similar to the ZFT ages at these sites, implying a rapid cooling from >250 to 180 °C at c. 90–95 Ma.

**Inverse modelling of low-temperature thermochronology data**

Only samples with over 20 dated grains and more than 50 track-length measurements were modelled. For each model (except that for sample HK11835; see discussion below), the computation was stopped after 100 ‘good’ paths (i.e. when both the model FT age and length distribution matched the measured FT age and length distribution with a level of goodness-of-fit of 0.5; see Ketcham 2005) were found. For samples HK11835 and HK12001 the absolute Cl content was used for the modelling, but for samples HK8758, HK11042 and HK12022 Dpar values were used and the c-axis correction was implemented for all samples (Ketcham 2005). The basic set of constraints for the intrusive rocks included the crystallization temperature and age, and the present-day temperature of 20 °C (Fig. 5a–c). Additionally, the ZFT age, if available, was used as a third constraint assuming a closure temperature at 250 ± 50 °C (Reiners & Brandon 2006). The inverse models operate only between 120 and 60 °C (i.e. apatite partial annealing...
Table 3. Detrital zircon and apatite fission-track ages for sedimentary rock samples

| Sample ID   | Lithology          | Stratigraphic Location | Coordinates                              | Mineral | Number of grains | $P_{\chi^2}$ (%) | $P$ (σ, ppm) | $P$ (χ2) % | $\rho_s$ (N (N − 1)) | $\rho_i$ (N (N − 1)) | Central age ± ρd (Ma) | $\sigma$ (ppm) | $\rho_i$ (N (N − 1)) | $\rho_s$ (N (N − 1)) |
|-------------|---------------------|------------------------|-------------------------------------------|---------|------------------|------------------|----------------|----------------|------------------------|------------------------|----------------------|-----------------|------------------|------------------|
| HK13325     | Sedimentary breccia | Cretaceous             | 22°33′18.72″, 114°17'27.37″               | Zircon  | 30               | 4.63 (4318)      | 142.0 (4630)   | 48.1 (1568) | 353         | 83.9 (0.00)            | 87.5 ± 18.6            | 4.63 (4318)      | 142.0 (4630)   | 48.1 (1568) | 353             |
| HK13326     | Sedimentary breccia | Cretaceous             | 22°33′05.45″, 114°15′38.16″               | Zircon  | 26               | 4.51 (4318)      | 130.6 (5938)   | 42.9 (723)  | 352         | 99.9 (0.00)            | 88.0 ± 19.4            | 4.51 (4318)      | 130.6 (5938)   | 42.9 (723)  | 352             |
| HK13326     | Sedimentary breccia | Cretaceous             | 22°33′05.45″, 114°15′38.16″               | Apatite | 10               | 13.91 (5023)     | 2.8 (86)       | 9.6 (298)   | 8           | 0.21 (0.52)            | 63.0 ± 26.2            | 13.91 (5023)     | 2.8 (86)       | 9.6 (298)   | 8               |
| HK13327     | Sandstone           | Cretaceous             | 22°30′40.27″, 114°20′01.90″               | Zircon  | 33               | 4.40 (4318)      | 129.7 (5887)   | 41.1 (798)  | 313         | 37.1 (0.00)            | 63.0 ± 26.2            | 4.40 (4318)      | 129.7 (5887)   | 41.1 (798)  | 313             |
| HK13329     | Sandstone           | Cretaceous             | 22°29′48.49″, 114°21′26.85″               | Zircon  | 49               | 4.21 (4318)      | 107.2 (2660)   | 40.2 (723)  | 306         | 92.75 (0.04)           | 73.1 ± 16.4            | 4.21 (4318)      | 107.2 (2660)   | 40.2 (723)  | 306             |
| HK13334     | Conglomerate        | Cretaceous             | 22°30′17.67″, 114°14′29.87″               | Zircon  | 31               | 4.12 (4318)      | 114.7 (2660)   | 40.2 (723)  | 298         | 93.13 (0.00)           | 76.1 ± 16.4            | 4.12 (4318)      | 114.7 (2660)   | 40.2 (723)  | 298             |
| HK13332     | Cherty siltstone    | Eocene                 | 22°32′59.52″, 114°16′14.07″               | Zircon  | 26               | 4.83 (5204)      | 120.3 (2935)   | 41.1 (798)  | 308         | 96.0 (0.00)            | 87.5 ± 18.6            | 4.83 (5204)      | 120.3 (2935)   | 41.1 (798)  | 308             |
| HK13332     | Sedimentary breccia | Cretaceous             | 22°32′27.44″, 114°25′′43.59″              | Zircon  | 28               | 4.68 (4318)      | 104.1 (1031)   | 37.0 (1113) | 260         | 100.0 (0.00)           | 87.6 ± 21.7            | 4.68 (4318)      | 104.1 (1031)   | 37.0 (1113) | 260             |

$\rho_s$ and $\rho_i$ represent sample spontaneous and induced track densities; $\rho_d$ is the standard track density; $\rho_d$, $\rho_s$ and $\rho_i$ are in units of $\times10^5$ tracks cm$^{-2}$. Ages were calculated using dosimeter glasses CN1 and CN5, where $\zeta_{CN1}$ (D.L.K.T.) = 129 ± 13 and $\zeta_{CN5}$ (A.C.) = 339 ± 5 (Hurford & Green 1983). $\chi^2$ is probability for obtaining $\chi^2$ value for $P_{n} = \text{number of crystals} - 1$. Central age is a modal age, weighted for different precisions of single crystals (Galbraith 1990).

Discussion

Provenance of post-magmatic sediments

The first-order observation of the detrital U–Pb age data is that most of the age populations from the four sedimentary units match closely the four major known magmatic episodes in Hong Kong (Davis et al. 1997; Campbell et al. 2007; Sewell et al. 2012b). Three of the five samples from Ping Chau, Cat O and Pat Sin Leng formations (HK11332, HK113326 and HK113344) also yield an age population at c. 152 Ma, corresponding to a period of magmatic activity identified by Sewell et al. (2012b), but with very limited outcrop areas. These data provide strong support to the idea that the source rocks of these units were the local volcanic–plutonic assemblages that were exposed at the time of deposition. The slight divergence of age components between the Cretaceous formations may reflect shifting of fluvial pathways and/or the existence or not of exposed strata in the relevant catchment(s). By Eocene time, the detrital zircons from the Ping Chau Formation also include a notable proportion of older Palaeozoic to Palaeoproterozoic grains, most probably sourced from reworked older sediments and basement rocks either from Hong Kong or mainland China, matching the ages of major orogenies there (Duan et al. 2011).

The youngest zircon population of c. 120 Ma from the Pat Sin Leng Formation, inferred to be the oldest post-volcanic stratigraphic unit (Lai et al. 1996; Sewell et al. 2000), provides a maximum age on the Cretaceous sediments. The presence of 120 Ma euhedral detrital zircons also suggests the presence of younger magmatic activity in this part of SE China after 140 Ma, although any surface expression of such volcanic or granitic rock is now missing in Hong Kong. Evidence, however, of a younger magmatic event is a granitic pluton, dated at 118.8 ± 2.0 Ma (LA-ICP-MS U–Pb zircon age), exposed in the neighbouring Shenzhen region less than 10 km from the northern boundary of Hong Kong (Shenzhen Geology Compiling Group 2009). The new data imply that the volcanic or granitic rocks were exposed at the surface, and were eroded into Cretaceous basins at some time soon after c. 120 Ma.

Post-eruption or post-emplacement reheating

The ZFT ages for the magmatic rocks are significantly younger than their eruption or emplacement ages. For volcanic suites, this indicates that the ZFT system has been reset by post-eruption thermal events. The granite ZFT ages, on the other hand, could indicate either a protracted post-crystallization cooling history or a reheating
event similar to that of the volcanic rocks and then cooling back again through the ZFT closure temperature prior to final exhumation. In the case of Hong Kong, we consider that the granites must have shared a similar tectonothermal history with the volcanic strata since the Early Cretaceous, and experienced the same post-eruption or post-emplacement reheating. This is based on the fact that the granites were shallowly emplaced within 3–4 km of the surface, and directly intruded the contemporaneous volcanic rocks; that is, they formed caldera-related volcanic–plutonic assemblages (Campbell & Sewell 1997; Sewell et al. 2012a). There is no structural evidence that the granites were ever separate from the volcanic rocks. The volcanic–plutonic rocks were adjacent and must have behaved as a single unit. Thus, the cooling of the granites (Fig. 5a–c) does not represent the original primary magmatic cooling, but corresponds to a secondary cooling phase after a post-emplacement heating event in association with the conjunct volcanic rocks. The primary magmatic cooling history of the granites is obscured by the post-emplacement heating event.

Further, since the Cretaceous sediments contain detrital zircons with U–Pb ages comparable with those of the local igneous rocks, we infer, therefore, that the volcanic–plutonic assemblages had ascended to the near-surface, and were being eroded during the middle to Late Cretaceous. This interpretation is also supported by the fact that the Cretaceous sediments were deposited on an erosion surface on the late Mesozoic volcanic strata. In addition, the reset detrital ZFT ages of the sediments are comparable with those of the igneous rocks, implying that the sediments and the volcanic–plutonic assemblages have probably all experienced a similar thermal history. These two lines of evidence imply that the volcanic–plutonic assemblages and the Cretaceous sequences have behaved in unison as a single package since middle to Late Cretaceous. They all underwent a post-emplacement or post-depositional heating event above the ZFT closure temperature (250° ± 50°C) and cooled through that temperature at c. 100–80 Ma. On the other hand, the detrital zircons from the Eocene Ping Chau Formation have not been reset and yield a range of detrital ZFT grain ages that is consistent with the reset ZFT ages of the local igneous and Cretaceous sedimentary rocks, serving to emphasize that the study area was exposed to temperatures of ≥250°C prior to the Eocene. Our results also imply that some of the Cretaceous sediments and the igneous rocks were returned to the surface, at least locally, by Eocene time.

**Source of heating: thermal event or re-burial heating?**

We have demonstrated that both the magmatic rocks and the Cretaceous sedimentary units have experienced a post-emplacement or post-depositional heating event sufficient to fully reset the ZFT system. Such heating could be the result of an elevated thermal gradient, or simply associated with burial. Assuming a typical continental geothermal gradient of 30°C km⁻¹, the rocks would have to be buried to depths exceeding 7–8 km to have fully reset the ZFT system solely by heating owing to burial. However, the Cretaceous sediments in Hong Kong, as well as in neighbouring Guangdong Province, were deposited in intermontane basins with restricted areal coverage. Nowhere in the region is there any preserved evidence for widespread and thick regional sedimentation during mid-Cretaceous time. In addition, if a 7–8 km thickness of sediments had been deposited regionally and subsequently eroded in the mid- to Late Cretaceous, an extraordinary tectonic mechanism, for which evidence is lacking, would be required for the rapid shift from a major depositional setting to an inversion of the Cretaceous strata. The rocks in general do not exhibit evidence for such a history (such as deformation, regional foliation and recrystallization), which would be expected had they undergone this hypothetical tectonic event.

Because it is unlikely that the heating was solely due to deep burial, we propose also that an enhanced geothermal gradient was present until c. 80 Ma. Two possibilities might account for such a gradient: (1) the Middle Jurassic to Early Cretaceous granitoids contain high concentrations of radioactive heat-generating elements (U, Th and K) that could cause a high heat flow, or (2) the continuing emplacement of younger intrusions. We consider that the first possibility is not valid because (a) the exposed granites in Hong Kong do not have enhanced concentrations of these elements (e.g. U ranges from c. 4 to 12 ppm in general; Sewell & Campbell 2001), and (b) if a higher heat flow owing to radiogenic sources was present in the mid-Cretaceous, such a heat flow should still be present. This is not the case, as regional heat flow in the Hong Kong area is at present <60–70 mW m⁻² (Tao & Shen 2008), which is comparable with the background values in continental crust of similar age and tectonic setting (Sclater et al. 1980; Rudnick & Fountain 1995). Therefore, the second explanation is more likely and implies that magmatism may have continued beneath Hong Kong until c. 100–80 Ma. These magmatic intrusions would have induced enhanced geothermal gradients, and driven geothermal or hydrothermal systems in rocks at shallower crustal levels above the plutons.

Evidence for hydrothermal activity at temperatures over 250°C is plentiful in Hong Kong. Petrographic examination of the 141 Ma High Island Tuff (Sewell et al. 2012a) revealed the presence of secondary epidote, adularia and sericite, typically associated with hydrothermal alteration (Browne 1978). Similar alteration mineral assemblages are evident in other volcanic and granitic rocks. The inferences from alteration mineral assemblages are supported by fluid inclusion micro-thermometry on two 140 Ma granites, which showed that the hydrothermal fluid temperature was over 290°C (D.L.K.T., unpubl. data). In addition, extensive quartz veins, indicative of hydrothermal events, and secondary epidote, which records formation temperatures of over 230–260°C (Bird & Spieler 2004), have been reported in the sedimentary rocks of the Pat Sin Leng, Port Island and Kat O formations (Lai et al. 1996).

We therefore interpret that these hydrothermal alterations were primarily driven by a heat pulse associated with magmatic intrusion or intrusions prior to the complete shutting down of Yanhsanian magmatism in SE China at c. 80 Ma (Zhou & Li 2000). Hydrothermally altered rocks in Hong Kong probably represent relict geothermal systems that were active during the late Mesozoic. Other fossil geothermal systems of similar age throughout the SE China Magmatic Belt have been related to ore deposition (e.g. Wang et al. 2012). Because plutons normally cool reasonably fast (of the order of less than 1 myr: Cathles 1977), the spread of the reset ZFT ages of over 20 myr (c. 100–80 Ma) implies that multiple heat pulses, corresponding to multiple magmatic intrusions, probably occurred.

**Post-volcanic thermal and tectonic evolution**

The FT thermal modelling (Fig. 5) reveals two main phases of cooling after the post-emplacement or post-eruptive heating event: a rapid cooling from mid-Cretaceous to Palaeocene–Eocene, followed by a decrease from then to the present. The phase of rapid cooling is inferred to primarily reflect geothermal readjustment after the cessation of the relevant intrusion-related hydrothermal system, with the combined effect of exhumation. As a result, we are unable to resolve the exhumation and erosion rate of the rock bodies inferred from the cooling rate determined
by our FT data prior to c. 60 Ma. We infer that the slower cooling (long-term averaged rate of the order of 1 °C m.yr⁻¹) since c. 60 Ma was a response to erosion-driven exhumation. This is equivalent to an averaged erosion rate of 0.03 mm a⁻¹, which is comparable with the average in the tectonically stable tropical and subtropical regions (e.g. Madagascar, Sri Lanka: Seward et al. 2004, Seward, unpubl. data). The estimated thickness of rock that has been eroded away since 60 Ma is c. 2–3 km. A similar, long period of slow exhumation and erosion has been identified in the Sierra Nevada, western USA since c. 60 Ma, during which less than c. 3 km of material have been removed (House et al. 1997, 1998; Cecili et al. 2006). The sedimentary rocks that are now exposed at the surface in Hong Kong probably represent the remnant of a thicker, eroded Cretaceous sequence. We infer that the slow exhumation since the early Cenozoic reflects an extended period of tectonic quiescence since the SE China had made the transition to a passive continental margin.

Our data and those of Yan et al. (2009) both show two main phases of cooling since the Cretaceous, although the specific timing of the phases is slightly different. The overall pattern indicates a period of rapid cooling followed by much slower cooling to the present day. Yan et al. (2009) argued that the cooling was related to uplift and denudation rather than any changes in heat flow rates, but our study suggests that a change in the geotherm was an important factor. Additionally, Yan et al. (2009) had undertaken analyses on only granitic rocks, and suggested that their FT ages and thermal models reflected the primary post-emplacement, prolonged cooling of the granites, without reference to a reheating event. Our data from Hong Kong extrusive and sedimentary rocks show, however, that this explanation cannot be correct. The timing of the change from rapid to slow cooling in Hong Kong and southern Guangdong within the APAZ is broadly similar at about 60 Ma (Fig. 5). Our FT data, however, are unable to verify the possible 34, 10 and 3–4 Ma tectonic events identified by Campbell & Sewell (2005), or the presence of higher cooling rates at c. 15–10 Ma as proposed by Yan et al. (2009). This is because these age periods are positioned outside the upper boundary of the APAZ (at c. 60 °C) in our modelled temperature-time paths where the inverse models of FT data are not reliable.

The presence of sodium-rich alteration minerals including aegirine and zeolite, which were interpreted as related to alkaline-rich hydrothermal alteration by Kemp et al. (1997), in the Eocene Ping Chau Formation indicates that a younger hydrothermal event, but of lower temperature, might have occurred. Our findings from the detrital ZFT analysis of this unit reveal that the ZFT system has not been reset since its deposition, implying that the Eocene hydrothermal event, if it occurred, must have reached temperatures below the ZFT closure temperature. The formation temperature of aegirine is debatable, although its presence has been recorded at a temperature of >130–160 °C in some modern hydrothermal systems (e.g. Yellowstone, USA: Browne 1978).

**Regional and global context**

Extensive Early Cretaceous granites (c. 120–80 Ma) are now exposed in Zhejiang and Fujian provinces, c. 250 km NE from Hong Kong, but are not exposed in the southern part of the SE China Magmatic Belt (Fig. 1). The conventional interpretation of this spatial distribution is that Yanshanian magmatism ceased in the southern part of the SE China Magmatic Belt after c. 140 Ma, and that subsequently the magmatic front shifted to the NE to Zhejiang and Fujian provinces (e.g. Zhou et al. 2006; Sewell et al. 2012a). We suggest, in contrast, that the Yanshanian magmatism may have continued until at least c. 100–80 Ma in Hong Kong and the nearby region. We infer that the temporal distribution of Yanshanian magmatic rocks essentially reflects the variation in the depth of crustal exhumation and erosion along the SE China Magmatic Belt, such that deeper crustal levels (and younger granites) are exposed in the NE region, but remain buried in the southern region (including Hong Kong).

The record of Yanshanian magmatism and Pacific margin evolution seen in Hong Kong has close parallels with other large-scale silicic igneous provinces in circum-Pacific regions. These include the Coast Mountains Batholith along the northern coast of British Columbia, Canada (180–100 Ma, Gehrels et al. 2009), the Sierra Nevada batholiths along the western coast of North America (210–88 Ma, Stern et al. 1981; Chot & Moore 1982), the Ibagué, Antioquia and Cordoba batholiths in Colombia (180–77 Ma, Villagomez et al. 2011), the Coastal Batholiths in central Chile (161–140 Ma, Gana & Tosdal 1996; Maksaev et al. 2006; Creixell et al. 2011), the silicic volcanic rocks in Graham Land, Patagonia and Antarctica Peninsula, Antarctica (188–153 Ma, Pankhurst et al. 2000; Riley et al. 2001, 2010; Leat et al. 2009) and the Median Batholith in South Island, New Zealand (170–100 Ma, Allibone & Tulloch 2004). This contemporaneity reinforces the link between the Yanshanian magmatism and the subduction of the palaeo-Pacific Plate, which formed a complete ring of circum-Pacific active plate margins during the late Mesozoic.

Many previous workers have linked the end of the Yanshanian events to rollback of the subducting slab (e.g. Shi & Li 2012), although the role of any larger-scale triggers (i.e. the interrelated circum-Pacific tectonic processes) has not been identified. All of the late Mesozoic silicic magmatic activity along the circum-Pacific regions, including the Yanshanian, came to an end at a similar time at c. 80 Ma. This coincident termination of magmatism around the Pacific coincided closely with the end of the Cretaceous Normal Superchron (c. 125–83 Ma; Gibbard et al. 2008; He et al. 2012) and other major plate reorganization events around the Pacific basin (e.g. Smith 2007; Seton et al. 2012). A major shift in the stress field around the entire Pacific margin plausibly links these tectonic processes.

In particular, for the SE China Magmatic Belt, the cessation of Yanshanian magmatism may have been triggered by the interaction between the Izanagi–Pacific spreading ridge and the eastern margin of the Pacific at c. 80 Ma. The approach of the oceanic ridge in Japan and Korea was constrained by heat flow anomalies and the cessation of granitic plutonism in the early Cenozoic (Muller et al. 2008; Seton et al. 2012). The collision of the Izanagi–Pacific spreading ridge along the SE China margin, which has never been examined in detail, possibly induced a change in subduction dynamics, leading to slab rollback. This event may also signify the inception of subduction of the juvenile Pacific Plate.

Following the withdrawal of subduction and tectonic plate rearrangement in the western Pacific, the proto-South China Sea began to open as a back-arc marginal basin between the SE China margin and the Borneo region at c. 80 Ma (Seton et al. 2012). Between this time and the beginning of full-scale opening of the South China Sea in the Oligocene, the SE China margin transformed from an active to a passive continental margin. The original forearc of the late Mesozoic magmatic arc was dismembered and displaced by crustal extension and rifting (Shi & Li 2012). Parts of the Yanshanian belt that originally lay outboard of Hong Kong (and the modern SE China seaboard) are found in the submerged continental shelf (Yan et al. 2014) or have been split off by spreading associated with the opening of the South China Sea. These granites and volcanic rocks with ages as young as 80–85 Ma are now found in Palawan, western Philippines and as micro-blocks in Nansha (e.g. Yan et al. 2010; Knittel 2011; Aurelio et al. 2013).
Conclusions

This study presents the first detailed investigation of the post-magmatic thermotectonic history of Hong Kong by employing low-temperature thermochronological techniques coupled with U–Pb dating of detrital zircons from post-volcanic sediments. The major findings are as follows.

(1) ZFT ages of the Mesozoic magmatic rocks in Hong Kong range between 140 and 60 Ma. Most are between c. 100 and 80 Ma, whereas the AFT ages range between 83 and 40 Ma, with two outliers at c. 28 and 23 Ma. The results show that both the ZFT and AFT systems have been reset, implying that the rock bodies experienced post-emplacement heating to temperature of ≥250 °C. We interpret the thermal effect to be a combined result of heating owing to burial plus the effects of an elevated geothermal gradient driven by continuing Yanshanian magmatism in the region.

(2) FT analysis and U–Pb dating of detrital zircons from the post-magmatic sedimentary sequences allow an improved understanding of the sediment provenance. The U–Pb age populations of detrital zircons from the Cretaceous sedimentary units match precisely the ages of the magmatic rocks exposed at present in Hong Kong, except for the earliest Pat Sing Leng Formation, which yielded an additional younger (120 Ma) detrital zircon population. This result confirms that the Cretaceous sediments were mostly sourced from local volcanic–plutonic assemblages. The 120 Ma detrital zircons from Pat Sing Leng Formation indicate (a) the occurrence of younger magmatic activity in Hong Kong or in nearby Guangdong Province, and (b) the maximum age of the Cretaceous basins in Hong Kong. The provenance of the Eocene sedimentary unit probably includes both the local volcanic sequences and recycled Cretaceous sediments, as well as the reworked sediments from Palaeozoic and older basement rocks, which are present both in Hong Kong and in mainland China.

(3) The late Mesozoic volcanic–plutonic assemblages and the Cretaceous sediments have behaved as a single package since at least 120 Ma. They were all heated to over 250 °C and began to cool again through the ZFT closure temperature by c. 100–80 Ma. We link the initial rapid cooling to both geothermal adjustment and unroofing by erosion. The slow cooling after 60 Ma was probably related solely to erosion-driven exhumation.

Previous interpretations of post-140 Ma geological history in Hong Kong were restrained by limited outcrop, and generally considered that little had happened since the Early Cretaceous. The results of this study demonstrate a dynamic post-magmatic history that was connected to evolution of the continental crust in SE China and its transition to a passive margin. We interpret that Yanshanian magmatic activity probably lasted longer than previously considered (e.g., Sewell et al. 2000; Zhou et al. 2006); that is, until at least c. 100–80 Ma in this part of SE China. The cessation of magmatism accompanied the onset of continental arc collapse that was related to the large-scale crustal extension in the region. The focus of crustal extension shifted elsewhere in the region in the early Eocene, prior to the opening of the South China Sea. By then, a passive continental margin had developed along the SE zone of China.

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