Ultrahigh-pressure (UHP) metamorphic rocks are often exposed at regional scale in orogenic belts. It is still an open question whether the rocks in UHP belts consist of a coherent part of subducted continental crust. This question is crucial for the interpretation of regional structural features of UHP belts, the subduction polarity of continental crust, and the exhumation process and mechanism of UHP rocks. As one of the largest UHP terranes, the Dabie–Sulu ultrahigh-pressure metamorphic belt has been the focus of extensive studies in the past 20 years. It is essential when studying the thermal structure of the terrane to properly interpret the larger regional structural features of UHP belts, consisting of a coherent part of subducted continental terrane. This is the result of different Fe3+/Fe2+ estimates of minerals and the re-equilibration of follow-up petrological studies (Carswell et al. 1997; Tabata et al. 1998; Li et al. 2005; Shi et al. 2014). Carswell et al. (1997) proposed that the Dabie eclogite-bearing terrane contains two structurally distinct tectonometamorphic units, the northern UHP eclogite and the southern high-pressure (HP) eclogite. There is a 0.8 GPa pressure gap between the units representing a 20 km crustal section that is missing. Additionally, the eclogites in the UHP unit do not display any regional northwards P–T gradient. Shi et al. (2014) concluded that the Dabie eclogite-bearing terrane is the result of regional crustal stacking, based on the observation of three tectonometamorphic units with different P–T conditions. However, P–T conditions recorded by the garnet–clinopyroxene Fe2+/Mg exchange thermometer show a large scatter in samples from the same area or outcrop. This is the result of different Fe3+/Fe2+ estimates of minerals and the re-equilibration of...
mineral compositions during extensive retrograde metamorphism (Carswell et al. 1997). The large scatter of P–T estimates has obscured the interpretation of regional thermal structure. Therefore, it is necessary to further investigate the thermal structure using other thermobarometers or phase-equilibrium calculations.

The Ti-in-zircon thermometer was calibrated by Watson and Ferry (Watson et al. 2006; Ferry & Watson 2007), and has been successfully used to estimate metamorphic temperatures (e.g. Baldwin et al. 2007; Harley 2008; Clark et al. 2009; Ewing et al. 2013; MacDonald et al. 2015). This thermometer has several merits: zircon is a widely occurring mineral, has a high closure temperature and a simple composition. The Ti-in-Zircon thermometer is particularly useful in HP and UHP environments because metamorphic rocks are typically saturated in TiO₂ as rutile, and SiO₂ as quartz or coesite. Unfortunately, the pressure effect on the Ti incorporation in zircon is yet to be precisely determined.

This study systematically sampled gneisses along an approximately 35 km-long north–south striking transect in the Dabie eclogite-bearing terrane, and used the Ti-in-zircon thermometer to estimate the metamorphic temperatures. Combining this data with the P–T conditions determined from eclogites in previous studies, we have attempted to further illuminate the thermal structure of the Dabie eclogite-bearing terrane and discuss possible implications for the regional-scale structure.

Regional geological setting and sample description

The Dabie Mountains are located in the eastern Qingshui–Dabie Triassic collision belt in central China (Li et al. 1989, 1993; Ames et al. 1993). The UHP metamorphic terrane is exposed in the east-central part of the Dabie Mountains, and is sandwiched between the North Dabie Magmatic Complex and the southern Susong Group (Fig. 1). The North Dabie Magmatic Complex predominantly consists of Cretaceous granitoids with various sized lenses composed of metamorphic rocks (Hacker et al. 1998). Mylonitic faults and intrusive boundaries define the contact between the North Dabie Magmatic Complex and overlying UHP rocks. The Susong Group was metamorphosed to amphibolite facies and structurally overlies the UHP rocks, separated by a mylonitic fault (Shi & Wang 2006; Shi et al. 2014).

On the basis of rock associations and early studies, the Dabie eclogite-bearing terrane can be divided into five units along its north–south profile (Fig. 1). Units I and II are composed of biotite gneiss, epidote–biotite gneiss, magnetite granitic gneiss and sparse eclogite lenses. Earlier studies showed that the P–T conditions for eclogite formation in the two units are different (Carswell et al. 1997; Shi & Wang 2006; Shi et al. 2014). Unit I eclogites formed in HP metamorphism, whereas Unit II eclogites experienced UHP metamorphism. Unit III mainly consists of epidote–biotite gneiss, epidote–biotite–phengite gneiss, marble, eclogite, mafic cumulative rocks and serpentine. Unit IV is composed of phengite granitic gneiss, biotite gneiss and eclogite lenses, and has similar rock types to units I and II. Unit V is made up of epidote–biotite gneiss, epidote–biotite–phengite gneiss, marble, eclogite, meta-cumulative rocks and serpentine, similar to Unit III in the rock association. Jadeite quartzite occurred only in Unit V (Xu et al. 1992; Liou et al. 1997).

All five units were sampled along a north–south trending profile (Fig. 1). The samples are granitic to tonalitic gneisses, consisting of feldspar, quartz, mica, epidote, chlorite, garnet, and accessory minerals such as titanite, magnetite, ilmenite, rutile, calcite and zircon (Table 1). In our previous study (Liu et al. 2001), coesite was identified as inclusions in zircon in half of the samples, suggesting that the gneisses had suffered coeval UHP or HP metamorphism, together with the eclogite lenses. Therefore, the present mineral assemblages of gneisses formed in the amphibolite facies during retrograde metamorphism. Rutile was identified in all the samples, suggesting that the rocks were saturated in TiO₂ and thus suitable for the application of the Ti-in-zircon thermometry. In some samples, rutile occurs as inclusion in zircon, which was identified by using laser Raman spectroscopy (Liu et al. 2001). In other samples, we identified rutile in the matrix of thin sections (Fig. 2), as a relict phase replaced by titanite and ilmenite.

Analytical methods

Zircons were separated using heavy-liquid and magnetic techniques, followed by hand picking under a binocular microscope. The zircon grains were mounted in epoxy resin disks and then polished. The cathodoluminescent (CL) images of the zircons were acquired before quantitative spot analyses, using a Nova NanoSEM 450 field emission scanning electron microscope (FESEM) at the Institute of Geology and Geophysics, Chinese Academy of Science (IGGGAS), at an accelerating voltage of 10 kV. For each sample, three to five zircon grains were selected for Ti, Y, Th and U analyses, and several grains were also selected for U–Pb age dating.

The Ti, Y, Th and U contents in the zircons were analysed using the Cameca Nano-SIMS 501 ion microprobe at the IGGGAS. Measurements were done using grain mode: first secondary ion images
are acquired by rastering a 20 × 20 µm area, and then spot analyses are carried out by deflecting the primary beam onto a selected area (c. 3 × 3 µm). For both ion imaging and spot analyses, an O⁻ beam (c. 100 pA) was optimized with the following configuration of the apertures (D₀–2 = 150 µm, D₁–3 = 200 µm) and the lens (L₀b = c. 6.6 kV, L₁ = c. 5.1 kV), corresponding to a typical probe size of 800 nm. The secondary ions were extracted from the sample surface by the co-axial immersion lens and focused onto a fixed-width analyser entrance slit (ES). About 50% of secondary ions were cut with ES-3 (relative to full transmission without slits) and the aperture slit (AS-1). The mass resolving power (MRP, the Cameca definition) at this condition is approximately 5000. Despite the generally high Y content of zircon, potential interference from Y is minimal because it only has one isotope, ⁸⁹Y. Although ⁴⁹Ti is only 5.5% of the total Ti, we selected ⁴⁹Ti for analyses because ⁴⁶Ti, ⁴⁷Ti and ⁴⁸Ti are interfered with by doubly charged ⁹²Zr (160 000 MRP is required to separate), ⁹⁴Zr (23 000 MRP) and ⁹⁶Zr (7200 MRP). Silicon was measured by ²⁹Si (4.67% of total Si). ²⁹Si can be separated from the interference of ²⁸SiH with an MRP higher than 3500. Before counting, the sample’s gold coating was removed by rastering a 25 × 25 µm area with a c. 500 pA O⁻ primary beam for 2 min. For ion imaging, a 100 pA O⁻ primary beam was used to raster a 20 × 20 µm area for 5 min. Images were processed using Cameca WinImage software. Compositional images were used when selecting analyses locations in the zircons to avoid nano-inclusions and cracks. The standard zircons M257 and 91 500 were repetitively measured by using grain mode (both ion imaging and spot analysis), and the results were used to calculate corresponding element contents. All errors are reported at the 1σ level, including the analytical uncertainties for individual spot measurements, the variations in

Fig. 1. Simplified geological map showing the division of units and sample locations.
| Samples   | Rock type                        | Mineral assemblage                                                | Mineral mode* | Rutile † | Coe in zircon‡ | Zircon shape and size§ | CL structures   |
|-----------|----------------------------------|-------------------------------------------------------------------|---------------|----------|---------------|------------------------|-----------------|
| 99db-187  | Epidote–biotite gneiss           | Pl(35) + Qtz(30) + Kfs(10) + Bt(10) + Ep(4) + Phn + Cal          | Identified    | Identified | 60–420 µm    | O-Eq, P(20%)           | C–M–R1–R2      |
| 99db-146  | Phengite gneiss                  | Pl(45) + Qtz(35) + Phn(10) + Mag(5) + Ep + Grt + Spn + Rt        | Identified    | –        | –             | 40–100 µm              | C–M–R          |
| 99db-119  | Magnetite gneiss                 | Pl(45) + Qtz(35) + Kfs(15) + Ky + Mag + Ep + Rt                  | Identified    | –        | –             | 50–150 µm              | C–M–R          |
| 99db-130  | Phengite gneiss                  | Pl(50) + Qtz(35) + Phn(10) + Mag + Bt + Chl + Ep + Rt            | Identified    | Identified| 30–190 µm    | O-Eq, P (30%)          | C–M–R1–R2      |
| 99db-107  | Biotite gneiss                   | Pl(55) + Qtz(30) + Bt(10) + Phn + Ep + Spn                       | Identified    | –        | 35–160 µm    | O-Eq, P (40%)          | C–M–R          |
| 99db-76   | Phengite gneiss                  | Pl(30) + Qtz(30) + Kfs(30) + Phn + Ep + Mag + Chl                | Identified    | –        | 30–170 µm    | P, O-Eq (10%)          | C–M–R          |
| 99db-19   | Epidote–mica gneiss              | Pl(45) + Qtz(25) + Bt(15) + Phn(5) + Ep(5) + Chl + Spn           | Identified    | Identified| 50–330 µm    | P, O-Eq (30%)          | C–M–R1–R2      |
| 99db-95   | Epidote–biotite gneiss           | Pl(45) + Qtz(25) + Kfs(10) + Bt(10) + Ep(5) + Phn + Rt + Spn     | Identified    | Identified| 30–360 µm    | O-Eq, P (25%)          | C–M–R1–R2      |
| 2001db-37 | Magnetite gneiss                 | Pl(25) + Qtz(45) + Kfs(20) + Mag(5) + Grt + Bt + Chl + Rt + Spn + Ilm | Identified    | Identified| 60–100 µm    | O-Eq, P (15%)          | C–DP–R1–R2     |
| 99db-22   | Magnetite–phengite gneiss        | Pl(40) + Qtz(30) + Kfs(20) + Mag (5) + Phn(3) + Spn              | Identified    | –        | 100–400 µm    | P, O-Eq (8%)           | C–M–R1–R2      |
| 2001db-33 | Epidote–biotite gneiss           | Pl(15) + Qtz(30) + Kfs(30) + Bt(10) + Ep(7) + Chl(3) + Phn + Mag + Rt + Spn + Ilm | Identified    | –        | 30–150 µm    | P, O-Eq (10%)          | C–M–R          |

Mineral abbreviations: Coe, coesite; Phn, phengite; the others are after Kretz (1983).
*The number in parentheses is the volume content of the mineral.
†See Figure 2, and Liu et al. (2001).
‡Liu et al. (2001).
§O-Eq: ovoid to equant, aspect ratios (length:width) <2:1; P, prismatic; aspect ratios from 2:1 to 4:1. The size of zircon was measured along the long axis.
the repeated measurements on the standards and the uncertainties of the recommended value of the standards.

Zircon U–Pb ages for six samples were dated using laser ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS) at the State Key Laboratory of Continental Dynamics at the Northwest University, China. The analytical spot diameter was set to 33 µm. The detailed instrumental parameters and analytical procedures are outlined in Yuan et al. (2008). Another sample (99DB-76) was dated using laser ablation-multiple-ion counting-inductively coupled plasma-mass spectrometry (LA-MIC-ICP-MS) at the Institute of Geology and Geophysics, Chinese Academy of Science. The analytical spot diameter was set to 5 µm. The detailed instrumental parameters and analytical procedures are described in Xie et al. (2017).

**Zircon morphology and internal structure**

Zircon external morphology and internal structure were investigated using microscope and CL imaging. Zircon external morphology and internal structure were investigated using microscope and CL imaging. (see Table 1). Most of the zircon grains are prismatic with aspect ratios (length:width) of between 2:1 and 4:1 in samples 2001db-33, 99db-22, 99db-76 and 99db-146. Prismatic and ovoid–equant forms (aspect ratio <2:1, rounded terminations or multifaceted) are equally abundant in samples 99db-107 and 99db-119. Ovoid–equant forms dominate zircon grains in samples 2001db-14, 2001db-37, 99db-19, 99db-95, 99db-130 and 99db-187.

Zircon grains from a sample usually have similar CL zones. Figure 3 displays the typical CL structures of the zircons from different samples. Except sample 2001db-37, the zircons can be separated into two groups. Type I has a core–mantle–rim (C–M–R) structure (99db-146, 99db-119, 99db-107, 99db-76 and 2001db-33), whereas type II displays a core–mantle–rim 1–rim 2 (C–M–R1–R2) structure (2001db-14, 99db-187, 99db-130, 99db-19, 99db-95 and 99db-22) (Fig. 3). In the zircons with a C–M–R structure, the cores show blurred oscillatory zoning to unzoned grey-bright luminescence, the mantles lack luminescence and the rims exhibit strong luminescence. Zircons in
samples 99db-146 and 99db-119 typically have very thin and discrete rims but some grains lack such rim. In zircons with C–M–R1–R2 structures, the R2 zone is the outermost bright rim; typically, very thin and discrete, some grains lack this rim. The R1 zone is the grey to dark-coloured inner rim. The M zone is grey-coloured in the CL images and sometimes contains a core with a different luminescence. In samples 99db-19 and 99db-130, most of the zircon cores show a blurred oscillatory zoning; but in samples 99db-95 and 99db-187, they are predominantly unzoned and grey to bright white in CL images. In sample 2001db-37, zircon grains consist of a dark rim (R1) and deep grey-coloured core (C) with dark patches (DP). In some grains, a strongly luminescent zone (R2) can be observed outside R1 (Fig. 3). In some instances, the R2 zone intrudes into the interior of the zircon grain. Zircon grains in this sample are rich in mineral inclusions and pores.

Zircon compositions

Zircons with a C–M–R structure

The compositional differences between different CL zones are striking (Fig. 4; see also the Supplementary material for results to this and the following subsections). The cores are characterized by high Th and Y contents, and Th/U ratios; the mantles have high U, and low Th and Y contents, and low Th/U
ratios, whereas the rims show the lowest Th, Y and U contents. The Y–Th/U diagrams highlight a compositional break between the C and M + R zones. The Th/U ratios in most of the rims are as low as those in the mantles. However, several analyses in the rims of zircons in samples 99DB-107 and 2001DB-33 have very low U contents (just several ppm) and, thus, relatively high Th/U ratios.

Zircons with a C–DP–R1–R2 structure

In sample 2001db-37, the cores have a distinctly different Th and Y content, and Th/U ratio from the DP, R1 and R2 zones (Fig. 4). The cores have a Th/U ratio >0.1, and a high Th and Y content, whereas the DP and R1 zones have a Th/U ratio <0.1, and a low Th and Y content (Fig. 4). The R2 zone has the lowest Th, U and Y content (Fig. 4).

Zircons with a C–M–R1–R2 structure

Different CL zones in C–M–R1–R2 zircons can be distinguished in Th–Y, Y–Th/U and U–Y diagrams (Fig. 4). The zircons from samples 99DB-19, 99DB-95, 99DB-130, 99DB-187 and 2001DB-14 have a high Y content and Th/U ratio in the core, variable Th/U ratio in the mantle, and high U content and low Th/U ratio in rim 1, and the lowest Th, U and Y content in rim 2. The Th/U ratio gradually decreases from the C to the M zones. In sample 99DB-95, an analysis with a very low U content in the R2 zone has a relatively high Th/U ratio. The zircons in sample 99DB-22 show high Th/U ratios in the cores; and low Th and Y contents, and Th/U ratios in the mantles and rims 1 and 2.

Zircon U–Pb ages, Th/U ratios and the origin of different zones

Zircon U–Pb ages in eight samples were analysed to determine the origin of different CL zones (Table 2; Fig. 5).

For the zircons with a C–M–R1–R2 structure in samples 99db-187, 99db-95, 99db-130 and 99db-19 (Fig. 5a), U–Pb dating shows that the cores have a discordant age, suggesting that they are relict protolith zircons that were recrystallized and modified in composition during metamorphism but did not reach complete re-equilibration. The mantles yield discordant ages in part, and concordant ages in part. The concordant ages plot into the UHP metamorphic time of 240–220 Ma (e.g. Li et al. 1989, 1993; Ames et al. 1993; Rowley et al. 1997; Hacker et al. 1998; Zheng et al. 2005; Liu et al. 2006). High Th/U ratios (>0.2) and discordant ages in parts of the mantle suggest that the mantle parts are also relict protolith zircons that were modified during the metamorphic process (Fig. 5a, c). Some of them reached complete re-equilibration in U–Pb age during metamorphism. Two analyses of the R1 zone with low Th/U ratios (<0.1) yield concordance ages plotting in the range of UHP metamorphic time, suggesting that the R1 zone grew during metamorphism. In sample 99db-22, the zircon cores with a Th/U ratio >0.1 are discordant in age, and the mantles with Th/U ratio <0.1 are concordant with an UHP age (Fig. 5b), indicating that the cores are relict magmatic zircons, whereas the mantles are metamorphic growth zircons. The zircons from sample 2001DB-37 were analysed in the R1 zone, and in the domains containing the C, DP and R1 zones. One analysis of the R1 zone yields an UHP age, suggesting that R1 is a metamorphic growth zircon. The other analyses give relatively concordant ages that are consistent with the UHP time when the Th/U ratio <0.1, and discordant ages far from the UHP age when the Th/U ratio is >0.1 (Fig. 5b, d). This also suggests that the DP and R1 zones are metamorphic growths, while the cores are inherited from the protolith. In sample 99DB-76 (Fig. 5c, f), the M and R zones yield a concordant age of about 230 Ma that plots in the range of UHP metamorphic time, whereas the core parts give a c. 730 Ma protolith age; indicating that the M and R zones are metamorphic growth zircons, whereas the cores are protolith zircons.

The U–Pb data show that metamorphic growth zircon has very low Th/U ratio, usually <0.1, whereas the Th/U ratio of relict protolith zircon is >0.1 and decreases with decreasing apparent age (Fig. 5c, d), which is a common feature for metamorphic recrystallized protolith zircon (Hoskin & Schaltegger 2003). The Th/U ratio of zircon may help us to use Nano-SIMS analytical results to determine zircon origins of different zones for the samples for which U–Pb dating is not analysed (including samples 99DB-107, 99DB-119, 99DB-146, 2001DB-14 and 2001DB-33) and for completely re-equilibrated protolith zircons. For the former case, the R1 and R2 zones in sample 2001DB-14, and the M and R zones in samples 99DB-107, 99DB-119, 99DB-146, 2001DB-14 and 2001DB-33, have low Th/U ratios, and are discontinuous in Th/U ratio with their relict protolith zircons, suggesting that these zones are metamorphic growth zircons. For the latter case, Figure 4 shows that some analyses in the C and M zones that are relict protolith zircons from samples 99DB-22, 2001DB-14 and 99DB-187 completely plot in the compositional fields of the metamorphic growth zircon on the Y and Th contents and Th/U ratio, suggesting that they had reached completely compositional re-equilibration during metamorphism, and can thus be used to estimate metamorphic temperature. However, determining the zircon origin by using its Th/U ratio appears not to be
Fig. 4. Zircon Th–Y, Y–Th/U and U–Y diagrams.
Fig. 4. Continued.
Table 2. Zircon analytical results of U–Pb ages and their Th/U ratios

| Spot No. | Th (ppm) | U (ppm) | Th/U | Isotopic ratio after common Pb corrected | Apparent age (Ma) | CL |
|----------|----------|---------|------|------------------------------------------|------------------|----|
| 99DB-187-4.1 | 362.47 | 533.47 | 0.679 | 0.06035, 0.001138, 0.01397, 0.08059, 0.00062 | 616, 32, 521, 8, 500, 4 | C |
| 99DB-187-4.2 | 267.20 | 473.03 | 0.565 | 0.05848, 0.00152, 0.56789, 0.01370, 0.07044, 0.00057 | 548, 38, 457, 9, 439, 3 | M |
| 99DB-187-5.1 | 308.58 | 374.17 | 0.825 | 0.06059, 0.00187, 0.60904, 0.01771, 0.07292, 0.00067 | 625, 47, 483, 11, 454, 4 | C |
| 99DB-187-5.2 | 60.50 | 191.89 | 0.315 | 0.05034, 0.00417, 0.25695, 0.02079, 0.03702, 0.00065 | 211, 150, 232, 17, 234, 4 | M |
| 99DB-95-10.1 | 27.57 | 90.57 | 0.304 | 0.04287, 0.00785, 0.21652, 0.03917, 0.03663, 0.00110 | 1197, 21, 599, 7, 454, 3 | C |
| 99DB-95-10.2 | 33.21 | 159.47 | 0.208 | 0.05517, 0.00429, 0.25648, 0.01951, 0.03372, 0.00055 | 850, 44, 515, 11, 442, 4 | C |
| 99DB-95-16.1 | 99.60 | 210.53 | 0.421 | 0.08000, 0.00146, 0.80419, 0.01287, 0.07290, 0.00049 | 736, 72, 655, 21, 632, 8 | M |
| 99DB-130.1.1 | 11.50 | 529.98 | 0.022 | 0.05164, 0.00102, 0.20270, 0.00359, 0.02846, 0.00018 | 270, 29, 187, 3, 181, 1 | R1 |
| 99DB-130.1.2 | 24.62 | 69.58 | 0.354 | 0.06738, 0.00199, 0.65995, 0.01847, 0.07104, 0.00059 | 655, 18, 364, 4, 320, 2 | M |
| 99DB-130.3 | 42.94 | 75.56 | 0.746 | 0.07936, 0.00187, 1.15505, 0.02514, 0.10555, 0.00082 | 1181, 31, 780, 12, 647, 5 | C |
| 99DB-130.4 | 12.19 | 23.77 | 0.513 | 0.06574, 0.00842, 0.29891, 0.04260, 0.03690, 0.00059 | 558, 326, 266, 33, 234, 4 | M |
| 99DB-130.4 | 36.82 | 121.10 | 0.304 | 0.03588, 0.00164, 0.36159, 0.01042, 0.04867, 0.00037 | 366, 51, 313, 8, 306, 2 | C |
| 99DB-19-17.1 | 302.09 | 265.17 | 1.139 | 0.06318, 0.00174, 0.91768, 0.02358, 0.10534, 0.00092 | 714, 40, 661, 12, 646, 5 | C |
| 99DB-19-17.2 | 31.76 | 70.70 | 0.469 | 0.06052, 0.000727, 0.33547, 0.03971, 0.04020, 0.00083 | 622, 270, 294, 30, 254, 5 | M |
| 99DB-19-3.1 | 78.60 | 64.10 | 1.226 | 0.06510, 0.000732, 0.79591, 0.08749, 0.08867, 0.00250 | 778, 186, 595, 49, 548, 15 | M |
| 99DB-19-3.2 | 32.33 | 35.13 | 0.920 | 0.06014, 0.01141, 0.29809, 0.05373, 0.03595, 0.00123 | 609, 352, 265, 44, 228, 8 | M |
| 99DB-22.1.1 | 112.83 | 171.85 | 0.657 | 0.06357, 0.00096, 0.79579, 0.00989, 0.09077, 0.00054 | 727, 16, 594, 6, 560, 3 | C |
| 99DB-22.1.2 | 1.79 | 107.45 | 0.017 | 0.05295, 0.00157, 0.23175, 0.00654, 0.03174, 0.00022 | 327, 52, 212, 5, 201, 1 | M |
| 99DB-22.12 | 1.44 | 41.94 | 0.034 | 0.05012, 0.00424, 0.25766, 0.02161, 0.03728, 0.00037 | 201, 171, 233, 17, 236, 2 | M |
| 99DB-22.15.1 | 130.33 | 203.56 | 0.640 | 0.06908, 0.00084, 0.68505, 0.00737, 0.08146, 0.00046 | 639, 14, 530, 4, 505, 3 | C |
| 99DB-22.19.0 | 252.84 | 269.39 | 0.939 | 0.06198, 0.00076, 0.79479, 0.00710, 0.09301, 0.00052 | 673, 10, 594, 4, 573, 3 | C |
| 99DB-22.19.2 | 1.45 | 50.30 | 0.029 | 0.05680, 0.00343, 0.27877, 0.01657, 0.03560, 0.00033 | 484, 116, 250, 13, 226, 2 | M |
| Sample | Ti-Zircon Temperature of Dabie HP–UHP Gneisses | C | M |
|--------|------------------------------------------------|----|----|
| 99DB-22.26.1 | 139.88 188.73 0.741 0.06348 0.00082 0.00082 0.84056 0.00082 0.09603 | 724 12 619 5 591 3 | C |
| 99DB-22.26.6 | 2.48 88.66 0.028 0.00507 0.00254 0.26293 0.01286 0.03754 | 0.00031 231 98 237 10 238 2 | M |
| 99DB-22.28.1 | 136.38 182.53 0.747 0.06259 0.00082 0.00082 0.85923 0.00089 0.09957 | 694 12 630 5 612 3 | C |
| 99DB-22.28.2 | 2.35 54.27 0.043 0.05291 0.03336 0.27237 0.01704 0.03733 | 0.00034 325 127 245 14 236 2 | M |
| 99DB-22.30.1 | 122.26 193.77 0.631 0.06137 0.00095 0.65896 0.00844 0.07786 | 0.00045 652 18 514 5 483 3 | C |
| 99DB-22.30.2 | 1.95 105.28 0.019 0.05120 0.00262 0.24425 0.01262 0.03460 | 0.00029 250 101 222 10 219 2 | M |
| 99DB-22.35 | 12.32 191.19 0.064 0.05086 0.00119 0.24472 0.00529 0.03489 | 0.00023 234 38 222 4 221 1 | M |
| 99DB-22.5.1 | 64.17 111.80 0.574 0.06124 0.00109 0.74242 0.01154 0.08791 | 0.00054 648 23 564 7 543 3 | C |
| 99DB-22.5.2 | 1.14 57.45 0.020 0.04669 0.00336 0.22185 0.01578 0.03446 | 0.00031 33 139 203 13 218 2 | M |
| 01DB-037-1.1 | 36.54 642.58 0.057 0.05929 0.00218 0.28728 0.01037 0.03514 | 0.00026 578 82 256 8 223 2 | M' |
| 01DB-037-1.2 | 146.75 1051.52 0.140 0.09457 0.00126 0.52783 0.00504 0.04047 | 0.00023 1520 10 430 3 256 1 | M' |
| 01DB-037-2 | 119.00 991.21 0.120 0.07080 0.00100 0.43918 0.00465 0.04498 | 0.00026 952 12 370 3 284 2 | M' |
| 01DB-037-3 | 74.35 1070.00 0.069 0.05732 0.00396 0.28231 0.01937 0.03572 | 0.00028 504 157 252 15 226 2 | M' |
| 01DB-037-4 | 119.76 1224.10 0.098 0.07187 0.00103 0.36109 0.00395 0.03644 | 0.00022 982 13 313 3 231 1 | M' |
| 01DB-037-5 | 81.05 1556.63 0.052 0.05187 0.00117 0.23861 0.00520 0.03336 | 0.00020 280 53 217 4 212 1 | R1 |
| 01DB-037-6 | 103.72 1589.28 0.065 0.05310 0.00142 0.29154 0.00761 0.03982 | 0.00025 333 62 260 6 252 2 | M' |
| 99DB-76-1a1 | 3.49 1730.79 0.002 0.05186 0.00080 0.25779 0.00472 0.03605 | 0.00050 280 40 233 4 228 3 | M |
| 99DB-76-1a2 | 13.63 1317.83 0.010 0.05010 0.00161 0.24960 0.00926 0.03613 | 0.00069 198 74 226 8 229 4 | M |
applied for the zircon with an extremely low U content (<5–6 ppm). Some analyses in R2 and R zones have a relatively high Th/U ratio in samples 2001DB-33, 99DB-76, 99DB-107 and 99DB-95 (Fig. 4; see also the Supplementary material). However, in terms of their low Y and Th content, and the latest formation time, they are obviously metamorphic growth zircons.

Results

The temperature estimates of Ti-in-zircon thermometer are displayed in Figure 6 (see also the Supplementary material), and a summary is given in Table 3.

The zircons that grew during metamorphism are the M and R zones in the zircons with a C–M–R
structure, the DP, R1 and R2 zones in the zircons from 2001DB-37, the R1, R2 and M zones of zircons in sample 99DB-22, and the R1 and R2 zones in the zircons of samples 99DB-19, 99DB-95, 99DB-130 and 99DB-187. Additionally, some of the relict protolith zircons, including the C zone in sample 99DB-22 and the M zones in samples 2001DB-14 and 99DB-187, reached complete re-equilibration during metamorphism and thus record significant metamorphic temperatures. These zircons are used to calculate metamorphic temperatures. Average metamorphic temperatures for these zones are given in Table 3. The standard deviations of the average temperatures are <50°C, reflecting that the Ti content in different zircon zones is relatively homogeneous.

In Unit I, the temperature from the M and R zones in sample 2001DB-33 ranges from 513 to 534°C. In Unit II, the zircons in sample 99DB-22 give the temperatures of 632, 589 and 552°C for the M, R1 and R2 zones, respectively, and three analyses of completely re-equilibrated cores give a temperature of 670°C. The zircons in sample 2001DB-37 give the temperatures of 568, 620 and 538°C for the DP, R1 and R2 zones, respectively. The zircons in samples 99DB-19 and 99DB-95 from Unit III yield temperatures of 765–784°C for the R1 zone and 575–648°C for the R2 zone. The zircons in five samples from Unit IV produce variable temperatures. The zircons in the southernmost sample 99DB-76 give 580°C for the M zone and 540°C for the R zone. Temperature increases towards the north, with zircons in sample 99DB-107 yielding temperatures of 676°C for the M zone and 622°C for the R zone. The zircons in sample 99DB-130 obtained 726°C for the R1 zone and 594°C for the R2 zone. The M and R zones of zircons in the two northermost samples 99DB-119 and 99DB-146

Fig. 6. Zircon temperature–Th/U ratio plots.
yield consistent temperature estimates from 578 to 616°C. In Unit V, the zircons give temperatures of 763, 753 and 720°C in sample 99DB-187, and 779, 644 and 658°C in sample 2001DB-14 for the completely re-equilibrated M, R1 and R2 zones, respectively.

Table 3. Summary of the temperature estimates from the Ti-in-zircon thermometer

| Samples      | C-M–R structure | C–M–R1–R2 structure | C–DP–R1–R2 structure |
|--------------|-----------------|---------------------|---------------------|
|              |                 |                     |                     |
| 99DB-146     | Th/U            | 0.887–1.735         | 0.005–0.059         | 0.006–0.023         |
|              | T°C             | 700–920(7)          | 578(13)             | 598(4)              |
|              | SD              | 22                  | 38                  |
| 99DB-119     | Th/U            | 0.668–1.308         | 0.010–0.017         | 0.008–0.072         |
|              | T°C             | 681–727(3)          | 585(6)              | 616(2)              |
|              | SD              | 25                  | 12                  |
| 99DB-107     | Th/U            | 0.389–1.125         | 0.021–0.043         | 0.021–0.527         |
|              | T°C             | 705–744(4)          | 676(9)              | 622(9)              |
|              | SD              | 31                  | 39                  |
| 99DB-76      | Th/U            | 0.465–0.829         | 0.004–0.010         | 0.003–0.256         |
|              | T°C             | 686–699(5)          | 580(6)              | 540(10)             |
|              | SD              | 37                  | 26                  |
| 2001DB-33    | Th/U            | 0.408–2.594         | 0.018–0.034         | 0.002–0.737         |
|              | T°C             | 685–826(8)          | 537(10)             | 514(10)             |
|              | SD              | 18                  | 32                  |

Note: the temperature estimates from the zircon core part were given as the range from minimum to maximum, and the temperatures of other zones as the average temperature.

The number in parentheses is the amount of the analytical spot. SD, standard deviation; C1, the core part in which the Th/U ratio and Y content plot in the fields of the R1 and R2 zones.

M1, the mantle part in which the Th/U ratio and Y content do not plot in the fields of the R1 and R2 zones.

M2, the mantle part in which the Th/U ratio and Y content plot in the fields of the R1 and R2 zones.

yield consistent temperature estimates from 578 to 616°C. In Unit V, the zircons give temperatures of 763, 753 and 720°C in sample 99DB-187, and 779, 644 and 658°C in sample 2001DB-14 for the completely re-equilibrated M, R1 and R2 zones, respectively.
Discussion

Pressures of re-equilibration protoliths and metamorphic growth zircons

The completely re-equilibrated relict and metamorphic growth zircons from different samples yield a large temperature range from about 500 to 800°C, with significant temperature differences in neighbouring units or samples. Currently, the Ti-in-zircon thermometer is only calibrated at 1 GPa, but the Ti uptake of zircon is pressure dependent (Watson et al. 2006; Ferry & Watson 2007; Ferriss et al. 2008; Tailby et al. 2011). Considering that the pressure dependence of the Ti-in-zircon thermometer may be as large as 50°C/GPa (Ferry & Watson 2007) or 100°C/GPa (Ferriss et al. 2008), it is possible that a large temperature change is due to a pressure effect.

The rocks we analysed experienced HP to UHP metamorphism, 1–3 GPa higher than the calibration pressure of 1 GPa. Such a pressure difference needs to be considered for our study. We can use inclusion phases and temperature estimates to roughly constrain the pressure of zircon formation.

For the zircons with a C–M–R1–R2 structure, coesite was identified in the re-equilibrated M and R1 zones in samples 99DB-19, 99DB-95, 99DB-130 and 99DB-187 by laser Raman spectroscopy (Liu et al. 2001) (Fig. 7, samples 99DB-19 and 99DB-130), suggesting that the recrystallization and compositional modification of the mantle and core parts, and the growth of R1 zones, occurred at UHP conditions. The completely re-equilibrated M and R1 zones in sample 99DB-187 yield the same temperature estimates (Table 3), further indicating that the completely re-equilibrated M and R1 zones formed at the same metamorphic condition. The R2 zones of zircons in these samples give obviously low temperature estimates, suggesting that they formed in the process of decreasing temperature, which most likely matches late retrograde amphibolites-facies metamorphism. For sample 2001DB-14, the completely re-equilibrated M zone yields the same

![Fig. 7. Zircon CL images and microphotographs showing the zircon internal structure and distribution of inclusion minerals.](image-url)
Table 4. Peak temperatures obtained from zircon, inferred pressure conditions and the $P$–$T$ conditions of eclogites at corresponding positions

| Units | This paper | Tabata et al. (1998) | Carswell et al. (1997) | Shi & Wang (2006) | Shi et al. (2014) | Others |
|-------|------------|----------------------|------------------------|-------------------|-------------------|--------|
|       | Samples    | $T$ (°C) | Pressure | $T$ (°C) at 3 GPa | $T$ (°C)/$P$ (GPa) | $T$ (°C)/$P$ (GPa) | $T$ (°C)/$P$ (GPa) |
|       | Ti-in-zircon | B(95) | GCP-GC* | GCP-GC† | GCP-GC/GCKS§ |
| Unit I | 2001DB-33 | 537 ± 18 | Qtz | 600–620/2.3 | 598 ± 58/1.90 ± 0.29 | 524 ± 52/2.46 ± 0.27$^\|$ | 477–647$^\|$ |
|       | 2001DB-37 | 620 ± 50 | Coe | 625 ± 86 | 709 ± 46/2.58 ± 0.34 | 602 ± 59/3.11 ± 0.2$^\|$ | 624–843$^\|$ |
| Unit II | 99DB-22 | 670 ± 19 | Coe | 616 ± 104–787 ± 87 | 690–880/3.2–3.8 | 846 ± 70/3.60 ± 0.36 | 748 ± 57/4.25 ± 0.44$^\|$ |
|       | 99DB-95 | 765 ± 31 | Coe | 784 ± 21 | 549 ± 43 | 720/2.8 | 617 ± 14/3.15 ± 0.15$^\|$ | 676 ± 16/3.28 ± 0.12$^\|$ |
| Unit III | 99DB-19 | 784 ± 21 | Coe | 847 ± 56 | 565 ± 24–614 ± 61 | 689 ± 13/3.34 ± 0.08$^\|$ | 670–730$^\|$ |
| Unit IV | 99DB-76 | 580 ± 37 | Qtz | 549 ± 43 | 720/2.8 | 617 ± 14/3.15 ± 0.15$^\|$ | 676 ± 16/3.28 ± 0.12$^\|$ |
|       | 99DB-107 | 676 ± 31 | Qtz | 549 ± 43 | 720/2.8 | 617 ± 14/3.15 ± 0.15$^\|$ | 676 ± 16/3.28 ± 0.12$^\|$ |
|       | 99DB-130 | 726 ± 39 | Coe | 847 ± 56 | 565 ± 24–614 ± 61 | 689 ± 13/3.34 ± 0.08$^\|$ | 670–730$^\|$ |
|       | 99DB-119 | 585 ± 22 | Qtz | 549 ± 43 | 720/2.8 | 617 ± 14/3.15 ± 0.15$^\|$ | 676 ± 16/3.28 ± 0.12$^\|$ |
|       | 99DB-146 | 598 ± 38 | Qtz | 549 ± 43 | 720/2.8 | 617 ± 14/3.15 ± 0.15$^\|$ | 676 ± 16/3.28 ± 0.12$^\|$ |
| Unit V | 99DB-187 | 753 ± 23 | Coe | 585 ± 46–786 ± 90 | 690–880/2.8–3.6 | 731 ± 32/3.75 ± 0.20$^\|$ | 780–890/4.4–4.8$^\|$ |
|       | 2001DB-14 | 777 ± 37 | Coe | 585 ± 46–786 ± 90 | 690–880/2.8–3.6 | 731 ± 32/3.75 ± 0.20$^\|$ | 780–890/4.4–4.8$^\|$ |

B(95), garnet (G)–clinopyroxene (C) thermometer, (Berman et al. 1995).

*aGCP-GC: GCP (Waters & Martin 1993) and GC (Powell 1985).

*bGCP-GC: GCP (Waters & Martin 1996) and GC (Krogh-Ravna 2000).

†GCP-GC: GCP (Ravna & Terry 2004) and GC (Krogh-Ravna 2000).

‡GCP-GCKS (K-kyanite, S-quartz/coesite) thermobarometer (Ravna & Terry 2004). The $P$–$T$ conditions from Shi et al. (2014) are the average results for units I, II, III and V, and the others are from single samples.

§Li et al. (2005), GC (Powell 1985) given at 3 GPa.

¶Schmid et al. (2003), GCP (Waters & Martin 1993) and GC (Krogh-Ravna 2000).
temperature as the completely re-equilibrated M and R1 zones of zircons in sample 99DB-187 from the same unit, suggesting that the completely re-equilibrated M zone of zircons in this sample formed in UHP conditions, whereas the R1 and R2 zones, with obviously low temperatures, should form during late retrograde amphibolites-facies metamorphism. The inclusion minerals in the zircons from sample 99DB-22 were well identified by laser Raman spectroscopy (Liu et al. 2001). The mantles contain coesite and jadeite inclusions, whereas the R1 zone has albite and quartz inclusions (Fig. 7, 99DB-22 grain 23). Therefore, the R1 zones in zircons in samples 99DB-22 and 2001DB-14 formed at the retrograde metamorphic stage, whereas it formed at the UHP stage in other samples (99DB-19, 99DB-95, 99DB-130 and 99DB-187).

The R1 zone in zircons with a C–DP–R1–R2 structure from sample 2001DB-37 gives the same temperature as the UHP mantles of zircons from sample 99DB-22 from the same unit. Thus, this temperature represents the temperature at the UHP stage. In the zircons with a C–M–R structure, quartz is the only SiO$_2$ phase as inclusion for all samples (Liu et al. 2001). Therefore, we infer that the M and R zones grew in the stability field of quartz.

When considering the pressure dependence of the Ti-in-zircon thermometer, the M and R zones in the zircons with a C–M–R structure need less of a temperature increase than the coesite-bearing zones in the zircons with a C–M–R1–R2 structure. The coesite-bearing zones in zircons usually give higher temperature values than quartz-bearing zones in zircons (Table 4). Therefore, significant temperature differences in neighbouring units or samples have nothing to do with the pressure dependence of the Ti-in-zircon thermometer.

The results from the Ti-in-zircon thermometer show that different zones of zircons from the completely re-equilibrated core to the metamorphic growth rims in sample 99DB-22 record decreasing temperatures (Table 3). Since the rock experienced a clockwise $P$–$T$ path and the completely re-equilibrated core gives a higher temperature estimate than the M zone, the recrystallization and compositional modification of the cores must have occurred at a higher pressure than the mantle growth (Fig. 8). Therefore, the zircons in this sample recorded a path of retrograde metamorphism with decreasing $P$–$T$ conditions (Fig. 8, path 1). Correcting the C and M zone temperature estimates by 50°C/$\text{GPa}$ (Ferry & Watson 2007) will result in a $P$–$T$ path with a striking temperature decrease (Fig. 8a, path 2). The completely re-equilibrated M, R1 and R2 zones of zircons with a C–M–R1–R2 structure in other samples (99DB-19, 99DB-95, 99DB-130, 99DB-187 and 2001DB-12) also reflect a decreasing temperature process. For the zircons with a C–M–R structure, the M and R zones in the three samples yield similar temperature results within their errors (99DB-119, 99DB-146 and 2001DB-37: Table 3), whereas in two other samples, metamorphic temperatures are decreasing from the M to the R zones (99DB-76 and 99DB-107: Table 3). From the R1 to the R2 zone in zircons with a C–DP–R1–R2 structure, metamorphic temperatures also decreased (Table 3). However, from DP growth to R1 growth, metamorphic temperatures increased from 568 to 620°C. This is the only case that may reflect a prograde process among the 12 samples collected.

The metamorphic temperature estimates from the Ti-in-zircon thermometer for the gneisses appear to represent the temperatures along the exhumation $P$–$T$ path. Their implication for pressure conditions may be further made clear by comparison with the temperature conditions from eclogites enclosed in those gneisses.

**Comparison with temperature estimates from the eclogites**

We consider the highest temperature estimate in each sample to represent the sample’s metamorphic peak temperature and, thus, use it to compare with the results determined in earlier studies. The results are shown in Table 4 and Figure 9.

The eclogites in units I (sample 2001DB-33), II (samples 99DB-22 and 2001DB-37), III (samples 99DB-19 and 99DB-95) and V (samples 99DB-187...
and 2001DB-14) have been extensively studied (e.g., Wang et al. 1992; Okay 1993; Carswell et al. 1997; Castelli et al. 1998; Tabata et al. 1998; Schmid et al. 2000, 2003; Xiao et al. 2000; Franz et al. 2001; Rolfo et al. 2004; Li et al. 2005; Shi & Wang 2006; Shi et al. 2014). The eclogites in Unit IV (samples 99DB-76, 99DB-107, 99DB-130, 99DB-119 and 99DB-146) were studied by Carswell et al. (1997), Tabata et al. (1998) and Shi et al. (2014). In these studies, temperature and pressure were estimated using the garnet–clinopyroxene Fe$^{2+}$/Mg exchange thermometer, garnet–clinopyroxene–kyanite–SiO$_2$ thermometer and garnet–clinopyroxene–phengite barometer, respectively. However, a significant scatter of temperature estimate is present for the eclogites from the same areas or, even, outcrop, which probably results from using different geothermometers, differing Fe$^{3+}$ estimates and varying degrees of mineral re-equilibration during extensive retrograde metamorphism (Carswell et al. 1997). Therefore, the estimates of peak temperature obtained from the Ti-in-zircon thermometer are only compared with the high-temperature estimates obtained from the eclogites at the same areas.

Our results coincide with the temperature estimates of eclogites from earlier studies (Fig. 9). Zircons from the Unit I gneiss yield the lowest metamorphic temperature estimates among all the samples. Unit I was widely considered a high-pressure eclogite zone with the lowest metamorphic temperature estimates of the whole eclogite-bearing zone (Wang et al. 1992; Okay 1993; Carswell et al. 1997; Li et al. 2005; Shi & Wang 2006; Shi et al. 2014). The eclogites from units III and V usually give high-temperature estimates of >700°C (Okay 1993; Zhang et al. 1995; Carswell et al. 1997; Tabata et al. 1998).
et al. 1998; Schmid et al. 2000, 2003; Rolfo et al. 2004; Li et al. 2005; Shi et al. 2014), which is consistent with our results. The temperature changes along the north–south-trending profile show that the temperature increases from 540 to 780°C from Unit I through Unit II and into Unit III (Fig. 9), then decreases from Unit III to Unit IV, and increases again from Unit IV to Unit V, which is consistent with the results of Tabata et al. (1998), Shi & Wang (2006) and Shi et al. (2014) (Fig. 9b–d). In Unit IV, our results from five samples show a large temperature variation, which also matches the results from Tabata et al. (1998) (Fig. 9b).

Therefore, the peak temperatures obtained from the zircons should represent the temperature conditions of eclogite-facies metamorphism. However, such consistency raises a question about the pressure dependence of the Ti-in-zircon thermometer. If the pressure dependence is 50°C/GPa (Ferry & Watson 2007), our temperature estimates should be increased by 70–200°C based on pressure estimates from the eclogites. This will cause temperature estimates of the Ti-in-zircon thermometer to be distinctly higher than the results from the eclogites. Therefore, the pressure effect on the Ti-in-zircon thermometer is likely to be less than expected.

**Tectonic implication**

The purpose of determining the thermal structure of the Dabie eclogite-bearing terrane is to reveal the regional structure of the terrane. The primary question is whether or not the terrane is a coherent subducted continental terrane or a stack of distinct tectonometamorphic slices with different equilibration conditions. The peak temperatures obtained from zircons along the profile perpendicular to the present amphibole-facies foliation do not display any consistent increasing or decreasing trend but, instead, temperature fluctuations (Fig. 9a).

This suggests that the Dabie eclogite-bearing terrane is a stack of distinct tectonometamorphic slices rather than a coherent northwards-subducted continental terrane, as proposed by Wang et al. (1992). The further question is how many slices the terrane contains.

The division of tectonic units in Figure 1 is based on rock associations and the $P$–$T$ conditions of the eclogites determined in earlier studies. Similar peak temperatures obtained from zircons from different samples in a unit and significant temperature differences between neighbouring units indicate that units I, II, III and V should be considered separate coherent tectonometamorphic units. On the other hand, it is unlikely that Unit IV is one coherent unit due to the presence of large temperature differences between neighbouring samples. The temperature differences from 50 to 204°C occur between neighbouring units and neighbouring samples within Unit IV (Fig. 10). It is unusual that such large temperature differences occur within distances of 1–4 km.

Correcting Fe$^{3+}$ estimates using Mossbaur spectroscopy and Micro-XANES (Schmid et al. 2003; Proyer et al. 2004) allows us to effectively estimate $P$–$T$ conditions using a garnet–omphacite–phengite assemblage. This geothermometer shows a 6–8°C km$^{-1}$ geothermal gradient during metamorphism. On the basis of this metamorphic geothermal gradient, large temperature differences within less than a 4 km distance imply that some crustal sections were eliminated from between samples or units by tectonic processes (Carswell et al. 1997). On the other hand, previous studies (e.g. Carswell et al. 1997; Shi et al. 2014) which used the garnet–omphacite–phengite thermobarometer show that different units formed under different metamorphic pressures. Carswell et al. (1997) determined that a pressure gap of 0.8 GPa is present between Unit I and Unit III (Table 4). Shi et al. (2014) showed that the average...
pressure gap between units is 0.5–1.0 GPa (Table 4). On the basis of an 8°C km\(^{-1}\) metamorphic geothermal gradient, the temperature differences from 50 to 204°C mean that the pressure gaps are from 0.2 to 0.8 GPa between neighbouring samples or units. This is largely consistent with the results of pressure estimates. Based on temperature estimates, seven or eight structurally distinct tectonometamorphic slices can be identified along the approximately 35 km-long profile (Fig. 10). Within Unit IV, the temperature difference between samples 99DB-107 and 99DB-130 is about 50°C. Considering the uncertainty of the Ti-in-zircon thermometer, this difference may not be significant. Thus, Unit IV may contain three or four tectonometamorphic slices (Fig. 10). Therefore, this study supports the hypothesis of previous studies (Lin et al. 2009; Shi et al. 2014) that the Dabie eclogite-bearing terrane can be understood as a stack of tectonometamorphic slices.

Faults separating the slices are not yet adequately identified along the profile. Carswell et al. (1997) observed a 50 m-wide south-dipping mylonitic zone between Unit I and Unit II, and thought that the ductile fault formed in late orogenic extensional collapse and was responsible for the elimination of at least 20 km of the crustal section. However, we do not rule out that stacking of distinct tectonometamorphic slices occurred during early eclogites-facies exhumation. Xue et al. (1996) identified three-stage deformation by detailed mapping in the Shima area, including syn-UHP south-vergent thrusting (D\(_1\)), north-vergent folding and top-to-the-north thrusting under amphibolites-facies conditions (D\(_2\)), and a third generation of folding at a shallower level (D\(_3\)). They inferred that south-vergent thrusting formed during early eclogites-facies exhumation. This may have caused stacking of distinct tectonometamorphic slices with different equilibration conditions. These early mylonitic zones of thrusting should be difficult to identify since they were recrystallized and folded during D\(_3\) deformation and metamorphism. The present structural feature of the Dabie eclogite-bearing terrane was mainly produced in the D\(_2\) deformational stage, which is characterized by low-angle, south- to SE-dipping foliation and NW–SE mineral stretching lineation defined by alignments of quartz, amphibole, biotite and epidote. The D\(_2\) north-vergent folding and top-to-the-north thrusting, of course, should be also another important reason for stacking of distinct tectonometamorphic slices.

Conclusions

The Ti-in-zircon thermometer was used to estimate the metamorphic temperatures of gneisses collected along an approximately 35 km north–south-trending profile in the Dabie eclogite-bearing terrane. The zircons with two–three zones of metamorphic growth and the completely re-equilibrated protolith zircons yielded two–four significant metamorphic temperatures. The decreasing temperatures from the cores to the rims suggest that the recrystallization and metamorphic growth occurred during an exhumation \(P–T\) path. The temperature values and temperature changes between neighbouring units or samples are identical to the results obtained from eclogites in earlier studies. Significant temperature differences among analysed gneisses show that the Dabie eclogite-bearing terrane consists of distinct tectonometamorphic slices with different equilibration conditions, within which seven or eight slices have been distinguished along the 35 km-long profile.

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