When did the ultrahigh-pressure rocks reach the surface? A \(^{207}\text{Pb}/^{206}\text{Pb}\) zircon, \(^{40}\text{Ar}/^{39}\text{Ar}\) white mica, Si-in-white mica, single-grain provenance study of Dabie Shan synorogenic foreland sediments

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Abstract

A single-grain study of synorogenic, Triassic–Jurassic quartz arenites of the eastern and southern forelands of the ultrahigh-pressure (UHP) Hong’an–Dabie orogen illustrates the utility and sensitivity of three provenance analysis techniques. \(^{40}\text{Ar}/^{39}\text{Ar}\) detrital white mica ages indicate that the Middle Triassic stratigraphic level reached \(\sim 400 \, ^\circ\text{C}\) at \(\sim 125\) Ma, probably heated by latest Jurassic–Early Cretaceous intrusions. Un-reheated single grains are as young as the depositional age of their sedimentary hosts and yield a Triassic–Jurassic age maximum that indicates a Hong’an–Dabie source, and some Paleozoic and Proterozoic ages that are mostly younger than \(^{207}\text{Pb}/^{206}\text{Pb}\) detrital zircon ages. Si contents of detrital white micas from the same samples range from 3.0 to 3.6 atoms pfu, with \(\sim 70\%\) of the micas having <3.3 Si atoms pfu. The grains with 3.3–3.6 Si atoms pfu probably originated from Hong’an–Dabie, indicating initial exposure of high-pressure (HP) and even ultrahigh-pressure rocks during the Middle Jurassic. The single-grain evaporation \(^{207}\text{Pb}/^{206}\text{Pb}\) ages of detrital zircons provide little indication that one of Earth’s most spectacular unroofing processes was active in the adjacent orogen. Proterozoic and Silurian zircon ages abound, indicating that the detritus originated from the northern margin of the Yangtze craton and regions to the southeast, in particular the South China fold belt and the Cathaysia block, but only a few Triassic zircon ages, indicating a Hong’an–Dabie source, were measured. The Triassic signal is smallest because growth of metamorphic zircon is rare and is mostly limited to rim formation on pre-existing zircons, and because neither a magmatic arc nor syn- to post-collisional magmatism have been mapped in the
Hong’an–Dabie orogen. Based on the white mica ages and the Si-in-white mica analysis, the Hong’an–Dabie orogen provided as much as 80% of the foreland detritus. The ages of the oldest micas suggest that exhumation of Hong’an–Dabie began at 240 ± 5 Ma and thus the average exhumation rates exceeded 2 mm/year during the Middle Triassic to Middle Jurassic.

Keywords: Single-grain $^{40}$Ar/$^{39}$Ar mica and $^{207}$Pb/$^{206}$Pb zircon dating; Detrital mica geobarometry; Synorogenic sediments; Provenance; South China craton; Dabie Shan

1. Introduction

Ideally, foreland basins record the unroofing history of their associated hinterland. The foreland basins along the Qin–Tongbai–Hong’an–Dabie Mountains of eastern China should record the exhumation, cooling and erosion of the upper parts of the high-pressure (HP) and ultrahigh-pressure (UHP) terranes of the Hong’an–Dabie orogen (Fig. 1). This orogen formed during northwest-directed subduction and subsequent exhumation of the South China Block from mantle depths in Triassic–Jurassic times (Hacker et al., 2000). A typical foreland basin is adjacent to the hinterland, but Nie et al. (1994) and Zhou and Graham (1996) suggested that detritus derived from the Hong’an–Dabie orogen was transported more than 1000 km to the west and deposited in the Triassic marine Songpan–Ganzi flysch basin.

Exhumation and cooling of the Hong’an–Dabie orogen were contemporaneous with Middle Triassic to Middle Jurassic siliciclastic deposition in the foreland and with the formation of Triassic metamorphic core complexes/basement uplifts in the Yangtze foreland fold-and-thrust belt (Figs. 1 and 2). These basement uplifts include the Hong’an–Dabie Shan (Hacker et al., 2000), Wudang Shan (Ratschbacher et al., in press), Lu Shan (Lin et al., 2000) Wugong Shan (Faure et al., 1996), Zhangbaling (Zhang et al., 2001b and unpublished results) and Dongling (this paper). Any of these domes/uplifts may have provided detritus for the foreland basins.

The Triassic–Jurassic Hong’an–Dabie orogen and its eastern and southern forelands were structurally and thermally reactivated in Cretaceous to Cenozoic times (e.g., Schmid et al., 1999; Ratschbacher et al., 2000; Lin et al., 2000, 2001). Grimmer et al. (2002) used apatite fission track thermochronology to determine the sedimentary provenance of the Yangtze foreland basins east and south of Dabie (their study overlaps regionally with this work), but found that both the orogen and its forelands were reheated during the Cretaceous to Cenozoic. The reheating annealed most pre-existing fission tracks, making provenance analysis using detrital apatites impossible.

The continental-scale Tan–Lu fault zone (Figs. 1 and 2), which bounds the Dabie Shan in the east, was a transtensional fault in the mid–late Cretaceous and Cenozoic and its activity is related to low-temperature (< 200 °C) cooling of its footwall and deposition on its hanging wall (Qianshan basin; Ratschbacher et al., 2000; Grimmer et al., 2002). Cooling at 90 to 60 Ma and at 55 to 35 Ma coincided with rifting marked by Late Cretaceous–Eocene red-bed deposition in eastern China. No study has documented earlier, in particular Triassic–Jurassic, tectono-thermal activity on the Tan–Lu fault zone.

We analyzed Triassic–Jurassic synorogenic foreland sedimentary rocks south and east of Dabie using four single-grain techniques: classical point counting, laser probe $^{40}$Ar/$^{39}$Ar geochronology of white mica, $^{207}$Pb/$^{206}$Pb thermal evaporation geochronology of zircon, and determination of the Si content of white mica by electron microprobe analysis. Our goals were to (a) assess the utility and sensitivity of these techniques for determining provenance of the eastern and southern Yangtze foreland deposits, (b) determine the extent of thermal overprinting in the Triassic to Cenozoic foreland sedimentary rocks, (c) estimate the degree to which detritus from multiple sources was mixed within the basins, and (d) determine when the UHP rocks in Hong’an–Dabie Shan were first exposed.
Fig. 1. Geologic overview, modified from Ratschbacher et al. (in press). Insets show principal units of the South and North China cratons (modified from Zhao et al., 2000) and histograms summarizing published zircon ages of different types (conventional U–Pb—concordant ages and lower and upper intercept ages, SHRIMP, and 207Pb/206Pb evaporation ages) from the South China Block (ABGMR, 1987; HBGMR, 1989; Kröner et al., 1993; Li et al., 1994, 1989, 1999; Li, 1996, 1999; Ames et al., 1996; Xue et al., 1996a,b, 1997; Rowley et al., 1997; Jian et al., 1998, 2000; Marayuma et al., 1998; Hacker et al., 1998, 2000; Qiu et al., 2000; Wang et al., 2001a; Tu et al., 2001; Chavagnac et al., 2001; Zhang et al., 2001a; Chen et al., 2001; Ayers et al., 2002; Zou et al., 2002) and the North China Block (Li et al., 1985; HBGMR, 1989; Kröner et al., 1993; Lerch et al., 1995; Song et al., 1996; Yu et al., 1996; Zhai et al., 1998; Wang et al., 2001b). Major tectonothersmal events are given by their approximate age ranges (see text for details).
Fig. 2. Geologic map of the eastern Dabie Shan and its southern and eastern forelands with sample locations. Composite stratigraphic section is adopted from the 1:50,000 geologic map of Yueshan area (ABGMR, 1974) and our own mapping (Gaitzsch, 2001). The stereoplot (equal area, lower hemisphere) from site J64 indicates N–S contraction by bedding-parallel slip and alignment of micas and pebbles. The structures of this site were tilted during mid-Cretaceous NW–SE transpressional deformation (Schmid et al., 1999; see restored structures in right-hand diagram). The stereoplot from site D569 indicates NNE–SSW transpressional deformation along dextral strike-slip faults, conjugate to the sinistral Tan–Lu parallel faults (see Grimmer et al., 2002 for regional data coverage). The site is close to a hinge of a NE plunging mid-Cretaceous syncline; bedding dips to the NE. Rotation axis is defined by trend, plunge, and rotation angle.
2. Chronology of basement rocks surrounding the Qinling–Hong’an–Dabie orogen

2.1. North China block

The North China Block is subdivided into western, eastern, and central blocks (Fig. 1; Zhao et al., 2000). The western and eastern blocks are Archean cratons linked by a suture formed by collision at \( \sim 1.8 \) Ga (Zhao et al., 2000). U/Pb zircon ages from the North China Block cluster at \( \sim 3.8, 3.3, 3.0, 2.5, \) and 1.7–1.8 Ga (Fig. 1; Yu et al., 1996; Song et al., 1996; Zhao et al., 2000). During Paleozoic times, the southern North China Block was an active margin. Characteristic U/Pb zircon ages are 470–490 and 390–410 Ma (Fig. 1; Lerch et al., 1995; Ratschbacher et al., in press).

2.2. South China Block

The South China Block is subdivided into a northern Yangtze Craton and a southern Cathaysia Block (Fig. 1). The oldest rocks are 2.9 Ga trondjhemite gneisses exposed on the northwestern Yangtze craton (Ames et al., 1996; Qiu et al., 2000). Paragneisses associated with the trondjhemite basement yield 2.9–3.2 Ga detrital zircons that suggest the existence of early Archean crust. U–Pb zircon ages of \( \sim 2.5 \) Ga (Yangtze), \( \sim 1.8 \) Ga (Cathaysia), and \( \sim 1.4 \) Ga (Cathaysia) indicate thermotectonic events (Fig. 1). The Yangtze–Cathaysia collision produced the 0.9–1.3 Ga Sibao orogeny, part of the worldwide Grenville event (Li et al., 2002). Large-scale Neoproterozoic rifting (0.7–0.8 Ga) within the South China Block began soon after Yangtze–Cathaysia amalgamation (Li et al., 1999). The ‘Caledonian’ South China fold belt within Cathaysia (Fig. 1) is characterized by 420–450 Ma folding and the intrusion of 400–440 Ma S-type granitoids (Li, 1998; Gilder et al., 1996; Chen and Jahn, 1998).

In the northern Yangtze Craton about 30 km south-east of Dabie, the Dongling dome constitutes the basement closest to but mostly unaffected by the Hong’an–Dabie orogeny (Fig. 2). This northeast-trending dome is a 15 × 5 km antiform exposing quartzite, garnet–mica–schist, gneiss, amphibolite and granitoids. The dome plunges below Paleozoic cover to the northeast and is covered by Jurassic volcaniclastic rocks to the southwest. Because the Dongling dome is an example for Yangtze basement and a likely source for the eastern Yangtze fold belt detritus, we dated Dongling zircons and white micas (see the sections below on \(^{207}\text{Pb}^{206}\text{Pb}\) and \(^{40}\text{Ar}^{39}\text{Ar}\) dating for analytical details). Three zircons from a high-grade paragneiss (D567 in Tables 1 and 2) yielded a \(^{207}\text{Pb}^{206}\text{Pb}\) weighted mean age (WMA) of 2372 ± 10 Ma (MSWD = 8.9, individual ages range from 2370 ± 2 to 2377 ± 10 Ma). Another 2299 ± 3 Ma zircon may belong to this group, but may have experienced Pb loss. Seven grains yielded a WMA of 2000 ± 9 Ma (MSWD = 49; individual ages range from 1971 ± 6 to 2016 ± 6 Ma), and two zircons yielded late Proterozoic ages of 692 ± 10 and 783 ± 7 Ma. All zircons from this sample are corroded. An \(^{40}\text{Ar}^{39}\text{Ar}\) age of 124.8 ± 1.2 Ma from synkinematic white micas from the Dongling paragneiss records cooling related to Cretaceous tectonic denudation (Grimmer et al., 2002).

3. Synorogenic sedimentary rocks within the Yangtze foreland fold-thrust belt

3.1. Stratigraphy, sediment composition, and deformation

Depositional ages for our samples were assigned using Chinese maps (ABGMR, 1975, 1987). Fig. 2 contains a composite stratigraphic section from the Yueshan area of the eastern Yangtze foreland fold-thrust belt, which summarizes the deposition of Middle Triassic to Middle Jurassic synorogenic foreland sedimentary rocks (ABGMR, 1974; Gaitzsch, 2001). Breitkreuz et al. (1994) presented the details of Lower to Middle Triassic deposition and demonstrated that unconformable deposition of Middle Triassic coarse carbonate breccias and Triassic–Jurassic siliciclastic sedimentary rocks marked the breakup of a pre-existing carbonate platform. Three major depositional sequences bounded by unconformities occur in the Middle Triassic to Middle Jurassic section (Fig. 2). The basal upper Middle Triassic (T2) Tongtonjian Formation comprises \( \sim 1700 \) m of uniform, micaeous siltstones. The Early to Middle Jurassic (J1–2) Xianshan Group contains more than 1400 m of mica-
rich siliciclastic rocks with coal intercalations. The Xianshan Group is divided into three fining-upward cycles, each starting with a basal conglomerate. Conglomerate lenses and channel structures indicate deposition in a fluvio-lacustrine environment. Upper Jurassic (J₃) volcanic and volcanoclastic rocks lie unconformably above the Lower–Middle Jurassic sediments.

Our sample localities are shown in Fig. 2. Quartz, white mica, and plagioclase occur in all samples. Biotite is rare to absent, and potassium feldspar is scarce except in samples Y126 and J55. All J₁–2 samples contain rutile, epidote, zircon, and apatite. Hornblende occurs only locally in samples J72, Y126 and Y129. Secondary iron and copper ores are common in the sandstones and suggest regional hydrothermal fluid activity. Metallic copper and copper minerals occur from grain-size (sample J64) to cm scale (samples Y124, Y129, Y134, Y136, J59, J64, J70, D569). Ore formation is probably related to 137–120 Ma magmatism within Hong’an–Dabie (Ratschbacher et al., 2000), when the major iron and copper deposits of the lower Yangtze river region formed (Xu, 1990).

Deposition of the Lower–Middle Jurassic siliciclastics occurred within an approximately north–south transpressive stress field (Fig. 2, localities J64 and D569 and Schmid et al., 1999). Syn-depositional soft-sediment deformation, commonly layer-parallel shearing (Fig. 2; station J64), is widespread in all Jurassic deposits. The entire sequence is gently folded with NE-trending axes in the eastern Yangtze foreland fold-and-thrust belt (Fig. 2; J64, D569) produced by 125–75 Ma transpression (Schmid et al., 1999).

3.2. Point-counting analysis

We inferred the sedimentary provenance of 10 thin sections of Jurassic sandstones from the southern and eastern Yangtze foreland fold-and-thrust belt by point counting (analytical procedures in Appendix A). Tables 1 and 3 and Figs. 2 and 3 contain sample descriptions, locations and the point count data. All samples show various amounts of secondary clay, sericite, calcite, and quartz growth. Each sample was assigned to one of three groups based on its composition (Fig. 3). Group 1 samples contain almost no feldspar and few lithic volcanic fragments (J57, J64, J70, J224, D569, Y124, Y134). A Group 2 sample contains significant feldspar, but no volcanic fragments (J55). Two Group 3 samples contain significant amounts of feldspar and volcanic fragments (J78, Y126). As an indicator of maturity, the QFL diagram distinguishes the majority of samples (Group 1 and Y126 of Group 3) from sample J55 (Group 2) and J78 (Group 3) that contain less stable minerals. Except for arkosic sample J55 that is suggestive of a stable craton source, all samples plot in the recycled orogen provenance. The distribution of sample compositions in the \( Q_mFL_t \) diagram supports the interpretations.

Fig. 3. Point-counting data from Dabie foreland sedimentary rocks displayed as ternary diagrams (QFL, QₘFLₜ, and \( L_mL_sL_v \)) and illustrating the relationship between composition of sandstones and their source areas.
deduced from the QFL plot. Including $Q_p$ in $L_4$ rather than in $Q$ demonstrates only a weak dependence of the overall composition from $Q_p$ (chert and microquartzite fragments). This is characteristic of mature samples and (excluding sample J55) indicates a recycled orogen affinity.

In contrast to the QFL and $Q_mFL_t$ plots, the $L_mL_cL_s$ diagram assigns greater weight to the nature of lithic fragments and their implications for possible source areas. Groups 1 and 2 are virtually indistinguishable from each other because feldspar is absent; they are suggestive of suture belts and rifted continental margins. The Group 3 samples plot in the mixed magmatic arc and rifted continental margin fields because of their high volcanic lithic fragment contents and comparatively small amounts of $L_m$ and $L_s$. In this group, sedimentary lithic fragments dominate, whereas metamorphic lithic fragments are rare (<3%). J78 and Y126 are the only samples that contain significant amounts of volcanic lithic fragments; these likely mark the onset of Late Jurassic volcanism. This suggestion is supported by the upper Middle Jurassic depositional age and the field observation that pyroclastics are intermixed with siliciclastics (sample J78). J55 is the only sample characterized by a pronounced quartzofeldspathic composition.

In summary, the point counting data indicate that possible source areas range from stable craton (QFL), transitional continental block and recycled orogen provenance ($Q_mFL_t$) to rifted continental margins or suture belts ($L_mL_cL_s$). The scarcity of lithic fragments and the lack of metamorphic lithic fragments points to a stable environment, probably cratonic basement or recycling of orogenic material.

### 3.3. Si-in-white mica microprobe analysis

The compositions of 112 detrital white mica grains from seven samples of Triassic, Jurassic, and Tertiary sedimentary rocks were analyzed by electron microprobe (Table 1, Appendix B, Fig. 4). A complete listing of the data is available at: http://www.elsevier.com/locate/chemgeo.

Middle Triassic sample J65 produced a bimodal distribution of white mica Si contents, with exclusively low Si (low: <3.3 Si atoms pfu; intermediate: 3.3–3.5 Si atoms pfu; high: >3.5 Si atoms pfu). In contrast, mica grains from Jurassic samples D569 and

### Table 1

Sample locations, lithology, stratigraphic age, sieve fractions and applied methods

<table>
<thead>
<tr>
<th>Samples</th>
<th>Lithology</th>
<th>N latitude</th>
<th>E longitude</th>
<th>Grain size [μm]</th>
<th>Stratigraphy</th>
<th>Methods</th>
<th>Minerals</th>
</tr>
</thead>
<tbody>
<tr>
<td>D567</td>
<td>Kfs–sil–crd–gneiss</td>
<td>30°33.800’</td>
<td>116°49.200’</td>
<td>100–300</td>
<td>Pt-basement</td>
<td>Pb, Ar, EMP</td>
<td>zr, mus</td>
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<tr>
<td>D569</td>
<td>Green, coarse sst</td>
<td>30°40.300’</td>
<td>116°53.035’</td>
<td>100–250</td>
<td>J$_{1-2}$xn$^1$</td>
<td>Pb, Ar, EMP, PC</td>
<td>zr, mus</td>
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<tr>
<td>J46</td>
<td>Purple, fine sst</td>
<td>30°55.160’</td>
<td>117°54.480’</td>
<td>63–125</td>
<td>T$_3$</td>
<td>Ar, EMP</td>
<td>mus</td>
</tr>
<tr>
<td>J55</td>
<td>Gray, fine sst</td>
<td>30°49.432’</td>
<td>117°16.334’</td>
<td>J$_{1-2}$xn$^2$</td>
<td>PC</td>
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<tr>
<td>J57</td>
<td>Gray, coarse sst</td>
<td>30°49.770’</td>
<td>117°16.430’</td>
<td>100–250</td>
<td>J$_{1-2}$xn$^3$</td>
<td>EMP, PC</td>
<td>mus</td>
</tr>
<tr>
<td>J59</td>
<td>Gray, coarse sst</td>
<td>30°49.650’</td>
<td>117°16.260’</td>
<td>100–250</td>
<td>J$_{1-2}$xn$^3$</td>
<td>Pb, EMP</td>
<td>mus</td>
</tr>
<tr>
<td>J64</td>
<td>Green, coarse grw</td>
<td>30°39.845’</td>
<td>116°53.680’</td>
<td>125–250</td>
<td>J$_{1-2}$xn$^1$</td>
<td>Pb, Ar, EMP, PC</td>
<td>mus</td>
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<td>J65</td>
<td>Ochre, fine sst</td>
<td>30°40.795’</td>
<td>116°58.920’</td>
<td>63–125</td>
<td>T$_3$</td>
<td>Ar, EMP</td>
<td>mus</td>
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<td>30°39.870’</td>
<td>116°59.295’</td>
<td>63–125</td>
<td>T$_3$</td>
<td>Pb</td>
<td>zr</td>
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<td>Gray, coarse sst</td>
<td>30°40.555’</td>
<td>116°50.230’</td>
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<td>J$_{1-2}$xn$^1$</td>
<td>EMP, PC</td>
<td>mus</td>
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<td>J72</td>
<td>Gray, coarse grw</td>
<td>30°39.720’</td>
<td>116°52.960’</td>
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<td>J$_{1-2}$xn$^2$</td>
<td>EMP</td>
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<td>30°24.325’</td>
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<td>J83</td>
<td>Red, fine cgl</td>
<td>30°32.575’</td>
<td>116°26.479’</td>
<td>125–250</td>
<td>E$_1$</td>
<td>Ar, EMP</td>
<td>mus</td>
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<tr>
<td>J224</td>
<td>Gray, coarse sst</td>
<td>30°40.332’</td>
<td>116°54.694’</td>
<td>J$_{1-2}$xn$^1$</td>
<td>PC</td>
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<td>Y124</td>
<td>Gray, coarse sst</td>
<td>30°19.146’</td>
<td>114°51.862’</td>
<td>100–315</td>
<td>J$_{1-2}$</td>
<td>Pb, EMP, PC</td>
<td>zr, mus</td>
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<td>Y126</td>
<td>Gray, coarse grw</td>
<td>30°16.545’</td>
<td>114°56.236’</td>
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<td>J$_2$</td>
<td>Ar, PC</td>
<td>kfs</td>
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<td>Gray, coarse grw</td>
<td>30°16.930’</td>
<td>114°59.360’</td>
<td>63–125</td>
<td>J$_2$</td>
<td>Pb, Ar, EMP</td>
<td>mus</td>
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<td>Gray, coarse sst</td>
<td>30°13.644’</td>
<td>115°03.461’</td>
<td>100–400</td>
<td>J$_2$</td>
<td>Pb, EMP, PC</td>
<td>mus</td>
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<td>Y136</td>
<td>Ochre, coarse sst</td>
<td>30°13.830’</td>
<td>115°03.140’</td>
<td>63–125</td>
<td>J$_1$</td>
<td>Ar, EMP</td>
<td>mus</td>
</tr>
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Sst: sandstone; grw: greywacke; cgl: conglomerate; Pt: Proterozoic (~2500–600 Ma); Middle Triassic (~240–230 Ma); J$_{1-2}$xn$^{1-3}$; Early–Middle Jurassic Xianshan Group 1,2,3 (~210–160 Ma); J$_1$: Early Jurassic; J$_2$: Middle Jurassic; E$_1$: Paleocene (~65–55 Ma). Pb: $^{207}$Pb/$^{206}$Pb single grain zircon dating; Ar: $^{40}$Ar/$^{39}$Ar single grain zircon dating; EMP: microprobe analysis (white mica); PC: Point-counting; zr: zircon; mus: white mica; kfs: potassium-feldspar; sil: sillimanite; crd: cordierite.
Y124 range from low Si ($n = 55$) to intermediate ($n = 21$) to high ($n = 3$). Jurassic samples Y129, J57 and J64 yielded exclusively low-Si micas with unimodal and bimodal distributions, but one spot on a low-Si grain yielded a content of 3.42 Si atoms pfu (J64–grain 7). Five samples (Y124, Y134, J70, J72, D569) contain intermediate to high Si micas with unimodal (J72, Y134) and polymodal (J70, Y124, D569) distributions. In some intermediate-Si grains, spots with >3.5 Si atoms pfu were identified (Y134–grain 9, Y124–grain 5). The highest Si content of all analyzed spots was 3.64 Si atoms pfu (D569–grain 1). The youngest sample, a Paleocene conglomeratic sandstone (J83, Fig. 2), contains principally intermediate (83%) to high (4%) Si phengite.

In summary, the Triassic rocks contain no high-Si white micas, the Jurassic rocks contain minor amounts of intermediate- and high-Si micas, and the Tertiary rocks contain dominantly intermediate- and high-Si micas.

### 3.4. $^{40}$Ar/$^{39}$Ar single-grain white-mica ages

One hundred and fifty six grains of detrital white mica from samples D569, J46, J64, J65, J83, Y129 and Y136 and four grains of detrital potassium feldspar from sample Y126 were dated by the $^{40}$Ar/$^{39}$Ar laser technique (Table 1, Appendix C, Figs. 5 and 6). The data are presented as age-probability spectra (ideograms; e.g., Deino and Potts, 1992; Stewart et
Fig. 5. Step-heating intragrain and intergrain comparisons of detrital white micas from the southern and eastern foreland of the Dabie Shan. 

Upper panel: an intragrain correlation plot showing the apparent ages of two heating steps of the same grain plotted against each other. Central panel: an intragrain correlation showing three to six heating steps from the same grain, plotted as consecutive release steps. The reproducibility of the ages within each set of heating experiments suggests that inhomogenously distributed excess $^{40}\text{Ar}$ is probably not present in these analyzed grains. Lower panel: an inter-grain correlation, showing coherent age groups of several step-heated micas in the same sample, indicates that inhomogenously distributed excess $^{40}\text{Ar}$ is unlikely in the detrital grains analyzed. See text for discussion. TFA, total fusion age (uncertainty reflects only analytical precision); WMA, weighted mean age; $J$, irradiation flux parameter.
A complete listing of the $^{40}$Ar/$^{39}$Ar data is available at: http://www.elsevier.com/locate/chemgeo.

$^{40}$Ar/$^{39}$Ar single-grain laser-fusion ages can be affected by excess $^{40}$Ar distributed either homogeneously or inhomogeneously within an individual grain (Scaillet, 1998). Progressively degassing a grain in multiple steps helps identify inhomogeneously distributed excess $^{40}$Ar; Fig. 5 shows that inhomogeneously distributed excess $^{40}$Ar is unlikely in the grains we analyzed. Homogeneously distributed excess $^{40}$Ar at the grain scale is difficult to rule out with confidence, but, when several detrital grains within a single sample (inter-grain correlation; Fig. 5) give the same age, the likelihood of excess $^{40}$Ar is low.

Upper Middle Triassic fine-grained sandstone J46 yielded a polymodal age distribution. The two youngest and smallest grains from this sample gave identical ages of about 120 Ma, suggestive of thermal resetting, in agreement with an apatite fission-track age of...
102 ± 4 Ma from the same sample (Grimmer et al., 2002). It is not clear if the other grains from this sample, whose ages range from Jurassic to late Proterozoic, were also partially reset. Sample J65, a second Middle Triassic fine-grained sandstone was also reset, and cooled below ~ 400 °C at about 130 Ma. Similarly, the potassium feldspars of sample Y126 were, with the exception of one grain dated at about 170 Ma, reset and cooled below ~ 200 °C at about 105 Ma. The apatite fission-track age of 65 ± 2 Ma from the same sample (Grimmer et al., 2002) suggests a cooling rate of 2.5 °C/Ma during the Late Cretaceous.

The Jurassic samples D569, Y129 and Y136 yielded a Triassic–Jurassic mode with a few older peaks in the age spectrum. Jurassic sample J64, sampled close to and displaying the same age components as D569, records a Late Devonian mode and Silurian, Triassic–Jurassic and late Proterozoic peaks. These Jurassic samples contain grains as young as their assigned depositional age. The three youngest grains from samples J64, D569, Y129, Y136 are 188 ± 5, 192 ± 3, 184 ± 3, and 187 ± 2 Ma, respectively; the WMA of the three youngest grains in all, non-reheated Jurassic samples is 188 ± 2 Ma (uppermost Early Jurassic). These youngest grains provide a maximum depositional age for the Jurassic sedimentary units and, taking a time span for erosion and transport into account, suggest that a Middle Jurassic (J2) rather than an Early to Middle Jurassic (J1–2) depositional age should be assigned to all our Jurassic samples. Micas from sample J83, a Paleocene red-bed conglomeratic sandstone, yielded age distributions comparable to those of the Jurassic samples D569, Y136, and Y129: a Triassic–Jurassic mode and a Devonian submode.

In summary, the 40Ar/39Ar detrital white micas contain Triassic–Jurassic, Devonian, Neoproterozoic, and Mesoproterozoic age components. In the eastern Yangtze foreland fold-and-thrust belt, Middle Triassic and deeper stratigraphic levels were heated to more than 400 °C.

3.5. 207Pb/206Pb single-grain zircon ages

We obtained 47 207Pb/206Pb single-zircon evaporation ages from basement and foreland samples (Tables 1 and 2, Appendix D, Fig. 7). Zircons with 206Pb/204Pb < 1000 were not considered for geological interpretation because they may yield geologically meaningless ages after common Pb correction. Only the Dongling basement (see Section 3.1) and the Jurassic sedimentary rocks contained zircons suitable for dating. Detrital zircons are difficult to discriminate by visible properties such as optical microscopy and SEM, because sedimentary transport processes often blur their original morphology. Most of the zircons we analyzed were pink, but a few were colorless or red. A mixture of well rounded, fragmented, poorly rounded, and euhedral zircons are present within the Jurassic sedimentary rocks.

No Archean zircons were found, but every sedimentary sample yielded at least one Proterozoic zircon. One multifaceted detrital zircon (sample Y124, grain 23) has a well-constrained age of 2358 ± 5 Ma. The next oldest age is slightly younger (2.26 Ga). Most zircons fall within a broad maximum from ~ 2.06 to ~ 1.70 Ga. Within this cluster three zircons (from samples Y134 and D569) yielded almost identical ages with a WMA of 1794 ± 1 Ma (MSWD = 0.23). Only one zircon age (1393 ± 7 Ma, sample D569) bridges the gap between this Paleoproterozoic cluster and the Neoproterozoic, for which we obtained three ages ranging from 690 to 800 Ma (samples D569, J59).

Phanerozoic zircons can be subdivided into Silurian and Permo-Triassic ages. Four zircons from two samples (D569, Y134) yielded a WMA of 430 ± 7 Ma (MSWD = 2.9). One Permian (sample J64, grain 1) and two Triassic (sample J59, grain 1; Y134, grain 21) zircons comprise the second Phanerozoic age group. One zircon evaporated in two steps (sample Y134, grain 4) yielded ages that overlap within error (Table 2). The age of the first step is suspect because of its high common lead content. This grain might be a Permian zircon with a Triassic rim; its older 279 ± 21 Ma age is indistinguishable from the 282 ± 5 Ma age of grain 1 of sample J64. The 217 ± 4 Ma zircon from Y134 (grain 21) contained abundant common lead (Table 2), which decreased together with the uncorrected 207Pb/206Pb ratio during the course of the analysis (90 scans); the corrected 207Pb/206Pb ratio remained constant. The 226 ± 10 Ma zircon (sample J59–grain 1) shows a constant 204Pb/206Pb ratio throughout the analysis, but a decrease of the 207Pb/206Pb ratio from ~ 0.0515 to ~ 0.0500, i.e. from latest Permian to Late Triassic.
Fig. 7. $^{207}\text{Pb}^{206}\text{Pb}$ evaporation zircon age data from Dongling and the Jurassic sedimentary rocks of the Dabie foreland displayed as probability plots.
### Table 2: \(^{207}\text{Pb}/^{206}\text{Pb}\) isotope data of single zircons

<table>
<thead>
<tr>
<th>Sample–zircon #</th>
<th>No. of scans</th>
<th>(^{207}\text{Pb}/^{206}\text{Pb})</th>
<th>(^{204}\text{Pb}/^{206}\text{Pb})</th>
<th>(^{207}\text{Pb}/^{206}\text{Pb})corr</th>
<th>(^{207}\text{Pb}/^{206}\text{Pb}) age [Ma]</th>
</tr>
</thead>
<tbody>
<tr>
<td>J59–zircon 1</td>
<td>90</td>
<td>0.054145</td>
<td>0.0002520</td>
<td>0.050674</td>
<td>226 ± 10</td>
</tr>
<tr>
<td>J59–zircon 3</td>
<td>90</td>
<td>0.062892</td>
<td>0.000375</td>
<td>0.062580</td>
<td>693 ± 3</td>
</tr>
<tr>
<td>J64–zircon 1</td>
<td>90</td>
<td>0.056738</td>
<td>0.0003440</td>
<td>0.051940</td>
<td>282 ± 5</td>
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<tr>
<td>J64–zircon 9</td>
<td>82</td>
<td>0.075007</td>
<td>0.0010100</td>
<td>0.060694</td>
<td>628 ± 4</td>
</tr>
<tr>
<td>D567–zircon 1</td>
<td>90</td>
<td>0.147730</td>
<td>0.0001800</td>
<td>0.145950</td>
<td>2299 ± 3</td>
</tr>
<tr>
<td>D567–zircon 5</td>
<td>90</td>
<td>0.153520</td>
<td>0.0001080</td>
<td>0.152700</td>
<td>2376 ± 2</td>
</tr>
<tr>
<td>D567–zircon 9</td>
<td>90</td>
<td>0.126030</td>
<td>0.0002100</td>
<td>0.123700</td>
<td>2010 ± 2</td>
</tr>
<tr>
<td>D567–zircon 10</td>
<td>90</td>
<td>0.062992</td>
<td>0.0000466</td>
<td>0.062555</td>
<td>692 ± 10</td>
</tr>
<tr>
<td>D567–zircon 11</td>
<td>36</td>
<td>0.121890</td>
<td>0.0005053</td>
<td>0.121660</td>
<td>1980 ± 20</td>
</tr>
<tr>
<td>D567–zircon 13</td>
<td>90</td>
<td>0.124280</td>
<td>0.0000438</td>
<td>0.124150</td>
<td>2016 ± 6</td>
</tr>
<tr>
<td>D567–zircon 14</td>
<td>90</td>
<td>0.068558</td>
<td>0.0002460</td>
<td>0.065290</td>
<td>783 ± 7</td>
</tr>
<tr>
<td>D567–zircon 16</td>
<td>90</td>
<td>0.122710</td>
<td>0.000375</td>
<td>0.122660</td>
<td>1995 ± 2</td>
</tr>
<tr>
<td>D567–zircon 19</td>
<td>20</td>
<td>0.099849</td>
<td>0.0014100</td>
<td>0.098044</td>
<td>1208 ± 35</td>
</tr>
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<td>D567–zircon 21</td>
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<tr>
<td>D567–zircon 23</td>
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<tr>
<td>D567–zircon 28</td>
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<td>0.0002920</td>
<td>0.152780</td>
<td>2377 ± 10</td>
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<td>0.152210</td>
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<tr>
<td>D567–zircon 35</td>
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<td>0.122480</td>
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<td>0.122220</td>
<td>1988 ± 11</td>
</tr>
<tr>
<td>D569–zircon 1</td>
<td>90</td>
<td>0.113730</td>
<td>0.0001360</td>
<td>0.112310</td>
<td>1837 ± 6</td>
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<tr>
<td>D569–zircon 3</td>
<td>90</td>
<td>0.127260</td>
<td>0.0000286</td>
<td>0.127340</td>
<td>2061 ± 4</td>
</tr>
<tr>
<td>D569–zircon 4</td>
<td>90</td>
<td>0.123590</td>
<td>0.0004499</td>
<td>0.123440</td>
<td>2006 ± 3</td>
</tr>
<tr>
<td>D569–zircon 5</td>
<td>63</td>
<td>0.059088</td>
<td>0.0002580</td>
<td>0.055566</td>
<td>435 ± 16</td>
</tr>
<tr>
<td>D569–zircon 6</td>
<td>90</td>
<td>0.063137</td>
<td>0.0001810</td>
<td>0.062773</td>
<td>700 ± 7</td>
</tr>
<tr>
<td>D569–zircon 7</td>
<td>90</td>
<td>0.109990</td>
<td>0.0005360</td>
<td>0.109670</td>
<td>1793 ± 5</td>
</tr>
<tr>
<td>D569–zircon 8</td>
<td>90</td>
<td>0.111050</td>
<td>0.0000111</td>
<td>0.111440</td>
<td>1823 ± 23</td>
</tr>
<tr>
<td>D569–zircon 9</td>
<td>90</td>
<td>0.115470</td>
<td>0.0000859</td>
<td>0.114730</td>
<td>1875 ± 5</td>
</tr>
<tr>
<td>D569–zircon 10</td>
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<td>0.088773</td>
<td>0.0000410</td>
<td>0.088525</td>
<td>1393 ± 7</td>
</tr>
<tr>
<td>D569–zircon 11</td>
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<td>0.0001786</td>
<td>0.113230</td>
<td>1851 ± 3</td>
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<td>D569–zircon 13</td>
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<td>0.0001720</td>
<td>0.055409</td>
<td>428 ± 8</td>
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<td>D569–zircon 14</td>
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<td>0.116030</td>
<td>0.0004790</td>
<td>0.109990</td>
<td>1799 ± 16</td>
</tr>
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<td>D569–zircon 15</td>
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<td>0.117160</td>
<td>0.0000889</td>
<td>0.116400</td>
<td>1901 ± 5</td>
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<td>D569–zircon 17</td>
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<td>1865 ± 16</td>
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<td>D569–zircon 19</td>
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<td>0.106490</td>
<td>0.0001610</td>
<td>0.104680</td>
<td>1708 ± 4</td>
</tr>
<tr>
<td>D569–zircon 20</td>
<td>90</td>
<td>0.108990</td>
<td>0.0002740</td>
<td>0.105660</td>
<td>1725 ± 12</td>
</tr>
<tr>
<td>D569–zircon 23</td>
<td>16</td>
<td>0.066205</td>
<td>0.0003130</td>
<td>0.066000</td>
<td>806 ± 38</td>
</tr>
<tr>
<td>Y124–zircon 23</td>
<td>90</td>
<td>0.154810</td>
<td>0.0003230</td>
<td>0.151150</td>
<td>2358 ± 5</td>
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<td>Y129–zircon 3</td>
<td>90</td>
<td>0.121950</td>
<td>0.0002570</td>
<td>0.119860</td>
<td>1940 ± 2</td>
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<tr>
<td>Y129–zircon 5</td>
<td>45</td>
<td>0.142860</td>
<td>0.0000545</td>
<td>0.142680</td>
<td>2259 ± 4</td>
</tr>
<tr>
<td>Y134–zircon 2</td>
<td>90</td>
<td>0.058039</td>
<td>0.0001990</td>
<td>0.055367</td>
<td>427 ± 4</td>
</tr>
<tr>
<td>Y134–zircon 3</td>
<td>90</td>
<td>0.082692</td>
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<td>0.053630</td>
<td>355 ± 11</td>
</tr>
<tr>
<td>Y134–zircon 4</td>
<td>45</td>
<td>0.083431</td>
<td>0.0002260</td>
<td>0.050868</td>
<td>234 ± 35</td>
</tr>
</tbody>
</table>

Zircons with high common lead contents \((^{204}\text{Pb}^{206}\text{Pb}>0.01)\) are italicized; they are not considered for geologic interpretation and are not plotted into Fig. 7.
WMA of these two Triassic zircons is \(218 \pm 4\) Ma (MSWD = 2.6). In summary, the \(^{207}\text{Pb}/^{206}\text{Pb}\) single zircons contain Permian–Triassic, Silurian, and mostly Neoproterozoic, and Paleoproterozoic age components.

4. Discussion

4.1. Thermal overprinting, sensitivity and suitability of geochronologic techniques to provenance analysis of the Triassic–Jurassic foreland sedimentary rocks

Geochronology presented here, in Ratschbacher et al. (2000) and in Grimmer et al. (2002) constrains the intensity and regional distribution of Cretaceous reheating of the synorogenic foreland sedimentary rocks. In the eastern Yangtze foreland fold-and-thrust belt, the nearly complete resetting of detrital white micas indicates that temperatures of \(~ 400\) °C were reached between 120 and 130 Ma at the depth of the fine-grained upper Middle Triassic sandstones; the upper Middle Triassic to Lower Jurassic stratigraphic levels had cooled to \(~ 100\) °C by about 100 Ma. This is similar to the Dongling basement, where a paragneiss (see above) had cooled to \(~ 400\) °C by about 125 Ma. In the southern Yangtze foreland fold-and-thrust belt, the basement plutons were at \(~ 300\)° by about 140 Ma. The Middle Jurassic stratigraphic level had reached \(~ 200\) °C by 105 Ma and \(~ 100\) °C by 65 Ma. Two interpretations might explain the widespread reheating: either the upper Middle Triassic–Lower Jurassic stratigraphic strata were buried more than 10 km during the Early Cretaceous, or the rocks were heated by latest Jurassic–Early Cretaceous plutons.

The Jurassic to Cretaceous strata east of Dabie may be locally 10 km thick (ABGMR, 1975, 1987), suggesting that low-grade burial metamorphism may have affected the Triassic section. A 158 ± 14 Ma apatite fission-track age from Middle Triassic sandstone of the eastern foreland (Grimmer et al., 2002) indicates regionally variable temperatures at the Triassic stratigraphic level and is consistent with the diagenetic nature of the foreland sedimentary rocks. This age predates the regional 145–120 Ma magmatism within Dabie and the Yangtze foreland fold-and-thrust belt (Ratschbacher et al., 2000) and indicates that burial may have been accomplished mostly during the deposition of the Middle Triassic through Middle Jurassic coarse clastic rocks. Their 3–5 km thickness (Fig. 2) implies temperatures of \(~ 100\) °C at the mid-Triassic level assuming a normal thermal gradient. Widespread higher temperatures seem less likely to have been caused by burial, and thus the widespread thermal event within the foreland sedimentary rocks is probably due to advective heating by major intrusions.

The intensity of Cretaceous reheating demonstrated herein and the apatite fission-track study of Grimmer et al. (2002) indicate that the zircon, sphene, and apatite fission-track geochronometers with closure temperatures of 250–100 °C are unreliable for regional provenance studies in the synorogenic foreland sediments of the eastern and southern Yangtze foreland fold-and-thrust belt (Fig. 8). The zircon and sphene chronometers might be reliable provenance indicators in the foreland north of the Hong’an–Dabie orogen where Cretaceous plutons are rare—at least in surface exposures. In contrast, the \(^{40}\text{Ar}/^{39}\text{Ar}\) ages from detrital micas provide reliable provenance information down to at least the Triassic level in the eastern and southern foreland belts.

Comparing the single-grain \(^{40}\text{Ar}/^{39}\text{Ar}\) white mica and \(^{207}\text{Pb}/^{206}\text{Pb}\) zircon ages yields a striking observation—from the detrital zircons alone, one would not infer that one of Earth’s most spectacular unroofing events, involving the exhumation of crustal rocks from mantle depths, was active only a few tens of kilometers away from the studied foreland deposits. We attribute this to the observation that within the Hong’an–Dabie orogen, the growth of metamorphic zircons was rare and was mostly limited to the growth of thin rims on pre-existing grains and that neither a magmatic arc nor syn- to post-collisional magmatism have been mapped. The subduction-zone thermal gradient of \(~ 5–10\) km\(^{-1}\) during collision and exhumation of the HP and UHP crust to midcrustal depths (Liou et al., 2000; Ratschbacher et al., 2000) obviously kept the orogen beneath the temperature for widespread zircon growth. We do not attribute the absence of a strong Hong’an–Dabie signal in the detrital zircon pattern to any specific climatic or catchment geometry effect because the detrital white micas clearly carry the Hong’an–Dabie orogen signal.
4.2. Evolution of the South China Block

Integrating published ages with our new ages from the Dongling dome and the foreland sedimentary rocks produces a better understanding of the Proterozoic evolution of the Yangtze Craton. The 2.3–2.4 Ga zircons found in the Dongling paragneiss and in the Jurassic sedimentary rocks of the southern Yangtze foreland fold-and-thrust belt are not known from the North and South China Blocks. These zircons may mark a previously unrecognized high-temperature event in the South China Block, they may be detrital (together with other zircons from the Dongling paragneiss) and derived from an already eroded Paleoproterozoic basement, or they may represent partial Pb loss from the ~2.5 Ga event in either the North China Block or South China Block (Fig. 1). U/Pb zircon crystallization ages in the 2.1–1.9 Ga range are distinctly of South China Block affinity and have been reported from the Dabie and Kongling areas (Rowley et al., 1997; Ayers et al., 2002; Qiu et al., 2000). Our 2.1–1.7 Ga detrital zircons are discordant, and the discordance may originate from this mid Paleoproterozoic event (Figs. 1 and 6), which is also manifested by the new ~2.0 Ga ages from the Dongling basement.

A distinct 1.9–2.1 Ga belt might have existed along the northern Yangtze Craton from the Qinling (Ratschbacher et al., in press) through Dabie into Korea, where ~2.0 Ga zircon ages have also been reported (Turek and Kim, 1996). Sinian rifting affected the entire South China Block and likely began at 830–820 Ma (Li et al., 1999). Neoproterozoic bimodal volcanics (800–650 Ma U–Pb zircon ages) and mafic dikes from central Korea to west-central China mark successful rifting along the (present) northern Yangtze margin (Lee et al., 1998; Rowley et al., 1997; Ames et al., 1996; Gao et al., 1990). Two zircons from the Dongling dome, five of our detrital zircons (810–690 Ma), and three ~600 Ma detrital white mica ages record this rifting event within the Dabie foreland sedimentary rocks.

4.3. Source areas

The Early to Middle Jurassic detrital white mica ages from the foreland sedimentary rocks indicate that
the Hong’an–Dabie orogen was the dominant source area for the foreland detritus, because no other region in eastern China could have provided white micas of this age. Almost all of our Triassic–Jurassic detrital white mica ages lie between 230 and 200 Ma, the major period of cooling though ~ 400 °C in Dabie (Hacker et al., 1998, 2000). Although Permian to Triassic mica ages have been reported from the Yangtze foreland fold-and-thrust belt (poorly dated Permo-Triassic intrusions, e.g., Chen and Jahn, 1998; Gilder et al., 1996), Dabie was likely also the source of the Triassic grains.

Pre-Triassic white mica and Paleozoic zircon ages indicate a source external to the Hong’an–Dabie orogen. Ordovician–Silurian inversion of the failed Neoproterozoic rift between Yangtze and Cathaysia may have formed the South China fold belt (Fig. 1), accompanied by Silurian magmatism (Fig. 1; Li, 1998). In particular, our ~ 430 Ma detrital zircon ages match the Silurian intrusion ages (400–440 Ma; e.g., Chen and Jahn, 1998). Our 390 to 310 Ma detrital mica ages may reflect cooling following this event, although Carboniferous white mica ages have rarely been reported from the South China fold belt (Charvet et al., 1999). Silurian–Devonian sedimentary rocks interpretable as South China fold-belt foreland basin deposits stretch into the Dabie foreland (Li, 1998). Thus, reworking of these Silurian–Devonian foreland sedimentary rocks and erosion of intrusive and metamorphic rocks from the SCFB most likely provided the majority of the Paleozoic detrital grains for the Jurassic Dabie foreland sedimentary rocks. The 0.7–0.8 Ga detrital zircons probably originated from Dabie, where most of the protoliths have Neoproterozoic zircon ages (e.g., Hacker et al., 1998).

In comparison to the abundant Triassic–Jurassic white micas (~ 66% of the total 40Ar/39Ar ages), 0.7–0.8 Ga Neoproterozoic zircons are scarce (~ 9%) within the foreland sediments; we would have expected a larger fraction of Neoproterozoic zircons, if the Dabie orogen were a major source (e.g., Hacker et al., 1998). Even highly discordant zircons may retain their original 207Pb/206Pb signature (e.g., Söderlund, 1996; Karabinos, 1997; Kröner and Willner, 1998), and thus the lack of 0.7–0.8 Ga zircons may result from incomplete resetting of older zircons by Neoproterozoic thermotectonic activity, as suggested by our ‘discordant suite’ (Fig. 7). The preponderance of 1.9–2.1 Ga zircons over 0.7–0.8 Ga zircons in the local source area, as observed in Dongling, may be a second reason. This is supported by the study of Ayers et al. (2002), which documented ~ 1.9 Ga instead of 0.7–0.8 Ga zircon cores in eastern Dabie. Outside the Hong’an–Dabie orogen, the South China fold belt and Cathaysia seemed to have been an important source area. Charvet et al. (1996) related 1.5–1.4 Ga ages to a poorly constrained mid-Proterozoic collision in Jiangnan (Figs. 1 and 9). The two detrital white mica ages of about 1.47 Ga and one 1.4 Ma detrital zircon age may fit into this period. In particular, our 1.8 Ga and the 1.4 Ga detrital zircons suggest a Cathaysian provenance (Li et al., 2002).

Reworking of older sediments may explain the dominance of highly mature quartz arenites, the presence of sedimentary lithic clasts and well-rounded zircons in the Dabie foreland. Possible source strata are Sinian and Silurian–Devonian. Preferred removal of less stable minerals due to a change in erosion/weathering behavior might be another mechanism. A climatic change is suggested by a change in the sedimentation in eastern China, where Upper Triassic evaporites and red-purple sabkha deposits were replaced by Jurassic ochre and gray-green fluvio-lacustrine siliciclastics. Red-bed deposition and evaporite precipitation resumed during the latest Jurassic–Early Cretaceous and lasted until the early Cenozoic. A similar climate change was documented in basins in western China. There, Triassic deposits record relatively arid climate, which was replaced by more humid conditions during the latest Triassic and Jurassic, when lacustrine deposits prevailed. Arid climatic conditions were established again in the Late Jurassic (Hendrix et al., 1992).

Based on the white mica ages and the Si-in-white mica analyses, the dominant source area was the Hong’an–Dabie orogen, which provided an estimated 50–80% of the foreland sediment. Areas south and east of the foreland probably provided the remainder (Fig. 9).

4.4. Implications for unroofing of the HP–UHP Hong’an–Dabie orogen

During prograde metamorphism, phengites show a quasi-linear increase in Si with increasing pressure and a moderate decrease in Si with increasing temper-
atures (Massonne and Szpurka, 1997). Although
about 80% of white micas from granitic rocks have
Si contents below 3.23 atoms pfu, some, particularly
those crystallized from Al-poor pegmatitic melts, are
high Si phengites with up to 3.53 Si atoms pfu (Zane
and Rizzo, 1999). A major argument against the
presence of pegmatitic high-Si phengites in our detri-
tal grain selection stems from the intra-grain spot
analysis, which indicates that at least some of the
micas have retrograde zoning (Fig. 4). Apart from
Hong’an–Dabie, high-pressure source rocks are
known from the ~ 950 Ma Jiangnan blueschists
(Charvet et al., 1996). As neither ~ 950 Ma white
micas nor zircons were identified in the Dabie fore-
land sediments, intermediate- and high-Si white micas
are likely to have been derived exclusively from the
Hong’an–Dabie orogen. There, the UHP units contain
phengites with 3.45–3.70 Si atoms pfu (Hacker et al.,
2000; Schmid et al., 2000) and the HP units have
phengites with 3.35–3.45 Si atoms pfu (Hacker et al.,
demonstrated that the Si content of Dabie white micas
decreases roughly with their ages from 3.48–3.30 Si
atoms pfu (230–220 Ma micas) to 3.35–3.10 Si
atoms pfu (210–180 Ma micas), but strongly retro-
gressed phengites also occur within HP–UHP rocks.
Thus, even the low-Si white micas may have been
derived from Dabie. In particular, the upper crustal
Foziling and Luzhenguang units of northern Dabie
contain Triassic low-Si white micas (Hacker et al.,

Fig. 9. Possible source areas for foreland detritus. Based on the white mica ages and the Si-in-white mica analysis, the principal source area was
the Hong’an–Dabie orogen, which provided ~ 50–80% to the foreland sediments; areas south(east) of the foreland provided the remainder.
A pre-Cretaceous restoration places these units on top of the HP/UHP rocks (Hacker et al., 1998) and as a result, these units were likely the first to be eroded during unroofing. The abundance of low-Si white micas in most of our foreland samples (J57, J64, J70, D569, Y129) demonstrates that mostly upper crustal material was eroded during the Jurassic. A contribution from retrogressed crustal material is suggested by the zoning measured in some micas (e.g., sample J64–grain 7, Fig. 4). Locally abundant intermediate- to high-Si phengites (J72, Y124, Y134 and the Paleocene sample J83, which shows a similar distribution) point to exposed HP and maybe UHP rocks in the source area.

The lack of Triassic zircons might be explained by methodological and statistical issues, as discussed in Appendix D, or simply to limited Triassic zircon growth in the source area. Concordant Triassic zircons from the Hong’an–Dabie orogen are rare. Ames et al. (1996) reported a few nearly concordant Triassic zircons from Dabie, but the majority are discordant or are characterized by Proterozoic cores with Triassic rims (Hacker et al., 1998). The Permian $^{207}$Pb/$^{206}$Pb ages might therefore represent discordant zircons that originated from Hong’an–Dabie. Less likely, they might have originated from the Permo-Triassic intrusions in southeast China (e.g., Chen and Jahn, 1998; Gilder et al., 1996). The few high-Si phengites, those retrogressed from high-Si phengites, and the two Triassic zircons may be evidence that the first exposure of the UHP rocks occurred in the Middle Jurassic (190–160 Ma). $^{40}$Ar/$^{39}$Ar potassium-feldspar diffusion domain modeling (Hacker et al., 2000) indicates that the presently exposed HP–UHP rocks had already cooled below 200 °C in the Middle Jurassic and were therefore already at shallow crustal levels during the Middle Jurassic, probably providing limited surface exposure.

The detrital white micas may also date the initiation of exhumation and thus provide an age limit for the peak of HP and UHP metamorphism in Hong’an–Dabie. Presumably, the oldest micas stem from the tectonostratigraphically highest units and were the first to cool during tectonic denudation (Hacker et al., 2000). Accordingly, exhumation would have begun at 240 ± 5 Ma, estimated by averaging the three oldest Triassic ages from each of our detrital mica samples not affected by Cretaceous re-heating. This age estimate agrees with ion microprobe U–(Th)–Pb ages (235–240 Ma) obtained from metamorphic growth rims on zircons from the HP and UHP units of Hong’an–Dabie that approximate peak pressure conditions during Triassic metamorphism (Hacker et al., 1998; Ayers et al., 2002). Thus, average exhumation rates of >2 mm/year prevailed between the Middle Triassic and the Middle Jurassic in the Hong’an–Dabie Shan.

5. Conclusions

A study of Triassic, Jurassic, and Paleocene synorogenic foreland sedimentary rocks south and east of the Hong’an–Dabie orogen using point-counting, single-grain $^{40}$Ar/$^{39}$Ar white mica and $^{207}$Pb/$^{206}$Pb zircon dating, and Si-in-white mica detrital white mica microprobe analysis provided first-order information on the utility and sensitivity of these techniques in regional provenance analysis, degree of thermal overprint, source areas, and time of initiation of exhumation and initial surface exposure of the ultrahigh pressure rocks in the Hong’an–Dabie Shan.

(1) In the eastern Dabie foreland Middle Triassic sandstones were heated to ~400 °C during the Early Cretaceous. These temperatures, exceeding those likely to have been caused by burial, are due to advective heating around intrusions. Our study and that of Grimmer et al. (2002) demonstrate that the zircon, sphene and apatite fission-track geochronometers are unreliable for regional provenance studies. $^{40}$Ar/$^{39}$Ar detrital white mica ages provide reliable provenance information in Jurassic and younger rocks.

(2) From $^{207}$Pb/$^{206}$Pb zircon ages alone, one would not infer that one of Earth’s most spectacular unroofing events had been active a few tens of kilometers from the studied foreland deposits. This is attributed to the observation that within the Hong’an–Dabie orogen, growth of metamorphic zircon is rare and is mostly limited to rim formation on pre-existing grains.

(3) Single-grain $^{40}$Ar/$^{39}$Ar ages are as young as the depositional age of their sedimentary hosts, have a Triassic–Jurassic maximum, which indicates a Hong’an–Dabie source, and some Paleozoic and Proterozoic
ages that are mostly younger than $^{207}\text{Pb}/^{206}\text{Pb}$ detrital zircon ages.

(4) The Early to Middle Jurassic detrital white-mica ages indicate that the Hong’an–Dabie orogen is the dominant source area, since no other region in eastern China provides white micas of this age; the Dabie is likely also the source of the Triassic grains.

(5) Reworking of Silurian–Devonian foreland sedimentary rocks and erosion of intrusive and metamorphic rocks from the South China fold belt most likely provided the majority of the Paleozoic detrital grains. The 0.7–0.8 Ga detrital zircons probably originated from Hong’an–Dabie, where most of the protoliths have Neoproterozoic zircon ages. 1.9–2.1 Ga detrital zircons likely were derived from eastern Dabie and/or the Dongling foreland massif, thus from a source close to the foreland deposits. Detrital white mica and zircon ages of about 1.45 Ga and in particular 1.8 and 1.4 Ga detrital zircons suggest a Cathaysian provenance.

(6) Intermediate- and high-Si phengites are likely to be derived exclusively from the Hong’an–Dabie orogen. Based on the white mica ages and the Si-in-white mica analysis, the Hong’an–Dabie orogen provided an estimated 50–80% of the foreland sediment.

(7) Few high-Si phengites, those retrogressed from high-Si phengites, and two Triassic zircons may be evidence for a first limited surface exposure of ultrahigh-pressure rocks in the Middle Jurassic (190–160 Ma).

(8) The detrital white micas may also date the initiation of exhumation and thus provide an age limit for the peak of ultrahigh-pressure metamorphism in Hong’an–Dabie. Assuming that the oldest micas stem from the tectonostatigraphically highest units, they first cooled during tectonic denudation. Accordingly, exhumation would have begun at 240 ± 5 Ma. Average exhumation rates thus were >2 mm/year during Middle Triassic to Middle Jurassic.

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Appendix A. Petrographic analysis by point-counting

A total of 550 to 580 grains from each of 10 thin sections were counted by using a half-automatic ELTINOR (Carl Zeiss) point counter. Point spacing was 0.2 to 0.4 mm on each line, depending on the grain size. Equally spaced lines all over the thin section were used. Grains were classified as monocrystalline quartz ($Q_m$), polycrystalline quartz ($Q_p$), plagioclase ($P$), potassium feldspar ($K$), and lithic fragment ($L$). The lithic fragments were further subdivided into metamorphic ($L_m$), volcanic ($L_v$), and sedimentary ($L_s$) lithic grains (Table 3); all others were counted as “miscellaneous” (mis).

Monocrystalline particles were counted as single mineral grains, whereas polycrystalline particles were counted as lithic fragments. Lithic fragments are, however, restricted to aphanitic microcrystalline materials smaller than 50 µm. Grains larger than 50 µm were counted as mineral grains, even if they apparently constituted lithic fragments. This modification affected only the counting of polycrystalline quartz fragments, which were counted as single grains if they exceeded 50 µm. The polycrystalline quartz fraction ($Q_p$) therefore represents principally chert or microquartzite fragments.

Quartz, plagioclase, and potassium feldspar were distinguished by the presence or absence of twinning, tartan twinning, perthitic texture and differing response to alteration. Because of the strong alteration, staining was not applied to the samples. Grains counted as quartz–mica tectonites ($L_m$) show a preferred planar fabric that was used to differentiate them from undeformed quartz–mica aggregates ($L_s$). Many polycrystalline quartz aggregates (counted as single grains) show distinctive deformation textures like undulose extinction, deformation lamellae, irregular grain boundaries, and the presence of subgrains. This indicates a tectonometamorphic source area (recycled orogen provenance). Volcanic lithic fragments ($L_v$) were rec-
ognized by the occurrence of microlithic or ophitic texture.

Appendix B. Si-in-white mica microprobe analysis

White micas were analyzed at the University of Freiberg with a JEOL JXA-8900R electron microprobe with five spectrometers at 15 kV acceleration voltage and a beam current of 20 nA. The diameter of the electron beam was 3 μm, counting times were set 20 s for Si, Al, Mg, Ca, Ba, and K, and 30 s for Fe, Na, Mn, and Ti. The standard sets of CAMECA and of MAC TM were used. In addition, a JEOL JSM-6400 Scanning Electron Microscope was used with an acceleration voltage of 20 kV and a beam current of 400 pA. During five iterations, the take-off angle was 40°, and peak measurements lasted 60 s. The white micas analyzed were solid solutions of the end-members muscovite, celadonite, paragonite, and margarite. From the celadonitic (Al[VI] + Al[IV] → (Mg, Fe)[VII] + Si[IV]), paragonitic (K[IV] → Na[IV]), and margaritic substitutions (K[IV], Si[IV] → Ca[VII], Al[IV]) of muscovite, the celadonitic substitution plays a major role during prograde HP–UHP metamorphism (e.g., Massonne and Szpurka, 1997). Within the muscovite–celadonite solid solution series phengites (KAl2/CoxMgx[Al1−x

$Si_3+xO_{10}(OH)_2$) comprise the intermediate members
The same grain size fractions were used for dating and microprobe analysis. The apparently least altered micas were probed from thin sections or from grain mounts. Electron beams were focused at the grain center, which we assumed to be least affected by alteration. Most grains were probed in more than one spot (randomly distributed between core and rim regions) to check for reproducibility and possible compositional zoning. Each spot measurement can be displayed as a normal (or Gaussian) probability distribution function. The sum of individual normal probability distributions is presented in cumulative probability diagrams (e.g., Stewart et al., 2001) that allow a better assessment of the data and their uncertainties.

Appendix C. $^{40}$Ar/$^{39}$Ar single grain laser dating

Our samples were crushed and the >63 μm size mica fraction extracted using standard mineral separation techniques. The separates were cleaned by ultrasound and later purified by handpicking. The separates were packaged in pure Cu foil, stacked in a pure SiO$_2$ vial together with foil-packaged neutron fluence monitors, and irradiated in three different irradiations (S38, S40 and S42) at the Oregon State University TRIGA reactor. We used Taylor Creek sanidine (USGS standard 85G003; Duffield and Dalrymple, 1990) with an assigned age of 27.92 Ma as a neutron fluence monitor.

Single mica grains were heated under UHV conditions using a Spectra-Physics continuous Ar-ion laser operating in TEM$_{00}$ mode. The evolved gas was purified during extraction by SAES ST-172 and ST-101 getters and a stainless steel cold finger and analyzed on a MAP 216 mass spectrometer fitted with a Baur–Signer ion source and a Johnston MM1 multiplier with a sensitivity of approximately $2 \times 10^{-14}$ mol/V. Analyses were corrected for system blanks and instrumental mass discrimination using the program EyeSoreCon, written by B.R. Hacker.

Appendix D. $^{207}$Pb/$^{206}$Pb single zircon dating

The $^{207}$Pb/$^{206}$Pb evaporation method is based on the studies of Kober (1986, 1987). Pb evaporation and analysis of single zircons was done at Freiberg using a FINIGAN MAT 262 mass spectrometer. Individual zircons were enclosed into a rhenium evaporation filament and initially heated to 1450 °C to release lead from metamict zones and other impurities. After this “cleaning” process, the zircon was heated in one step to 1600 °C. The lead (and other) isotopes were evaporated and collected on a second rhenium ionization filament. From the ionization filament the lead was ionized at $\sim$ 1200°C (1180–1260 °C). Data acquisition by magnetic peak switching of the mass sequence $^{206}$Pb (4 s), $^{207}$Pb (4 s), $^{204}$Pb (8 s) used 4 s intermass delay time and ion counting. Baseline counts were determined on mass 204.5 every five blocks. Data acquisition comprises up to 10 blocks of 10 mass scans, depending mostly on the stability of the ion beam. The $^{207}$Pb/$^{206}$Pb ages were calculated from the measured $^{207}$Pb/$^{206}$Pb and $^{204}$Pb/$^{206}$Pb ratios with the following corrections: (1) common lead correction after Stacey and Kramers (1975), (2) a specific mass-spectrometer calibration factor (mass bias) calculated from measurements with two zircon standards. This mass bias (0.36 ± 0.22% amu) included the thermal fractionation of Pb and the mass bias of the ion counter. A standard (NBS 981), which contains all stable lead isotopes, was used for mass spectrometer calibration. Standard zircons were routinely measured in order to check for the reproducibility of the zircon ages.

The evaporation method requires a minimum of accumulated lead for a detectable and constant ion beam. The content of radiogenic lead in zircon crystals depends on the primary uranium content, age (i.e., time for lead accumulation), and the geologic history (possible lead loss events). Zircon grains >100 μm were selected preferentially. Thus, possible age groups restricted to grains <100 μm have escaped observation. To get sufficiently high lead beam intensities, entire zircon grains were evaporated. If a zircon had an older core and a younger rim and was evaporated in one step, the evaporation method would have resulted in a mixed age. This mixed age gives a minimum age for the core and a maximum age for the rim. The evaporation method often produces ages which are in agreement with independently determined ages—even from highly discordant zircons (e.g., Kröner and Willner, 1998; Karabinos, 1997). Söderlund (1996) showed that the evaporation method
produced more precise zircon ages than conventional U–Pb dating. However, as no information about the degree of discordance is available, uncertainties about the $^{207}\text{Pb}/^{206}\text{Pb}$ ages remain. Cogenetic discordant zircons may appear as a suite of different $^{207}\text{Pb}/^{206}\text{Pb}$ ages, which are all younger than their crystallization age (e.g., Stewart et al., 2001; Kröner and Willner, 1998). However, if several zircons have statistically indistinguishable ages they are commonly regarded as true crystallization ages; it is unlikely that different grains have lost exactly the same amount of lead.

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