

Deciphering igneous and metamorphic events in high-grade rocks of the Wilmington Complex, Delaware: Morphology, cathodoluminescence and backscattered electron zoning, and SHRIMP U-Pb geochronology of zircon and monazite

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ABSTRACT

High-grade rocks of the Wilmington Complex, northern Delaware and adjacent Maryland and Pennsylvania, contain morphologically complex zircons that formed through both igneous and metamorphic processes during the development of an island-arc complex and suturing of the arc to Laurentia. The arc complex has been divided into several members, the protoliths of which include both intrusive and extrusive rocks. Metasedimentary rocks are interlayered with the complex and are believed to be the infrastructure upon which the arc was built.

In the Wilmington Complex rocks, both igneous and metamorphic zircons occur as elongate and equant forms. Chemical zoning, shown by cathodoluminescence (CL), includes both concentric, oscillatory patterns, indicative of igneous origin, and patchwork and sector patterns, suggestive of metamorphic growth. Metamorphic monazites are

chemically homogeneous, or show oscillatory or spotted chemical zoning in backscattered electron images.

U-Pb geochronology by sensitive high resolution ion microprobe (SHRIMP) was used to date complexly zoned zircon and monazite. All but one member of the Wilmington Complex crystallized in the Ordovician between ca. 475 and 485 Ma; these rocks were intruded by a suite of gabbro-to-granite plutonic rocks at 434 ± 5 Ma. Detrital zircons in metavolcanic and metasedimentary units were derived predominantly from 0.9 to 1.4 Ga (Grenvillian) basement, presumably of Laurentian origin. Amphibolite to granulite facies metamorphism of the Wilmington Complex, recorded by ages of metamorphic zircon (428 ± 4 and 432 ± 6 Ma) and monazite (429 ± 2 and 426 ± 3 Ma), occurred contemporaneously with emplacement of the younger plutonic rocks. On the basis of varying CL zoning patterns and external morphologies, metamorphic zircons formed by different processes (presumably controlled by rock chemistry) at slightly different times

and temperatures during prograde metamorphism. In addition, at least three other thermal episodes are recorded by monazite growth at 447 ± 4 , 411 ± 3 , and 398 ± 3 Ma.

Keywords: Wilmington Complex, U-Pb geochronology, SHRIMP, zircon, monazite.

INTRODUCTION

The Wilmington Complex of northern Delaware and southeastern Pennsylvania is located within the central part of the Appalachian orogen. It is composed primarily of mafic to felsic magmatic rocks, some of which have been metamorphosed to amphibolite or granulite grade, which provide an excellent opportunity to understand the roles of early Paleozoic arc magmatism and orogeny in building the Appalachian mountain belt. Although the Wilmington Complex has been the subject of numerous investigations for more than one hundred years (e.g., Chester, 1890; Bascom et al., 1909; Ward, 1959; Woodruff and Thompson, 1972, 1975), interpretations of structural and petrologic

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studies (cf. Wagner and Srogi, 1987; Srogi et al., 1993; Srogi and Lutz, 1997) have been hindered by the lack of a precise and accurate geochronologic framework upon which tectonic models can be developed. Previous determinations of emplacement ages have been of uncertain accuracy due to complicating factors, such as inheritance of Proterozoic material and multiple subsequent episodes of Paleozoic metamorphism. If a temporal context can successfully constrain the magmatic and tectonic history of the Wilmington Complex, the resulting models may be applicable to other regions within the Appalachians or other arc terranes, leading to a better understanding of the development of complex arc-continent systems.

The bedrock geologic map of Delaware and adjacent Pennsylvania (Schenck et al., 2000; Plank et al., 2000) identifies several new units within the Wilmington Complex, including both metavolcanic and metaplutonic rocks. As part of the compilation for the map, U-Pb geochronologic studies were initiated to understand both the timing of igneous and metamorphic events that affected these rocks and, by implication, the tectonic conditions responsible for the development of the complex. This report uses the stratigraphic nomenclature of Schenck et al. (2000) and Plank et al. (2000) for consistency with the bedrock geologic map of Delaware. Zircons in several metaigneous rocks were dated to determine protolith ages and timing of subsequent deformation. Because the rocks of the Wilmington Complex have undergone amphibolite to granulite facies metamorphism, many of these zircons contain multiple age components that formed during both igneous and metamorphic events. Early attempts to date these rocks by conventional U-Pb geochronology (Grauert and Wagner, 1974, 1975) yielded equivocal results because of these complex zircons. Our study utilized the high spatial resolution of *in situ* analyses by the sensitive high resolution ion microprobe (SHRIMP) to decipher the age components within individual zircons. In addition, the ages of metamorphic monazite were determined from two samples. We use the newly obtained age data, combined with interpretations of zircon and monazite morphology and chemical zoning patterns, plus geologic mapping and geochemical studies, to develop a comprehensive geologic history for the Wilmington Complex.

REGIONAL GEOLOGY

The Wilmington Complex (Figs. 1 and 2) was originally defined by Ward (1959) to include igneous and metamorphic rocks from Cecil County, Maryland, across northern Delaware, to Chester, Pennsylvania. On the bases of new

detailed geologic mapping (Schenck et al., 2000), major and trace element geochemical data and interpretations of protoliths of units (Plank et al., 2001), and new U-Pb zircon ages, the rock units within the Wilmington Complex have been redefined (Plank et al., 2000). Two rock units now comprise Ward's (1959) "banded gneiss": the Rockford Park Gneiss is composed of interlayered mafic and felsic orthogneiss interpreted to be of volcanic origin, whereas the Brandywine Blue Gneiss, predominantly felsic orthogneiss containing thin, discontinuous mafic layers, is of possible plutonic origin. Units mapped by Ward (1959) as "amphibolite facies gneiss" are now defined as the Barley Mill Gneiss, Christianstead Gneiss, Faulkland Gneiss, Mill Creek Metagabbro, Montchanin Metagabbro, and Windy Hills Gneiss (Plank et al., 2000; Schenck et al., 2000). The Faulkland Gneiss and the Windy Hills Gneiss are interpreted to be volcanic in origin. The Christianstead Gneiss is a granodioritic pluton that intrudes the Windy Hills Gneiss. Following Ward (1959) and subsequent workers, intrusive rocks of the composite Arden Plutonic Supersuite also are included within the Wilmington Complex. These include the Ardentown Granitic Suite, Perkins Run Gabbro-norite Suite, and Biotite Tonalite (Plank et al., 2000). Rocks of the Arden Plutonic Supersuite are less deformed and recrystallized than layered gneissic rocks of the Wilmington Complex, and they intrude the Brandywine Blue Gneiss and the Wissahickon Formation.

The Wissahickon Formation, first named by Bascom (1902), is an extensive sequence of pelitic and psammitic gneisses interlayered with amphibolites, which borders the Wilmington Complex to the west, north, and east. For a detailed summary of the history of work on the Wissahickon Formation in Maryland and Delaware, see Schenck (1997). Structural data (Plank et al., 2000) demonstrate that the Wilmington Complex and Wissahickon Formation shared a common deformational history beginning in the Ordovician. The Arden Plutonic Supersuite intrudes the Wissahickon Formation on the northwest (Fig. 2). Based on these associations, we conclude that the Wissahickon Formation is older than the Arden Plutonic Supersuite.

Many workers have suggested correlation of the Wilmington Complex with the James Run Formation in Maryland. Originally defined by Southwick and Fisher (1967) as interlayered quartz amphibolite and biotite-quartz plagioclase gneiss, the James Run Formation was redefined by Higgins (1971, 1972) to include all of the metavolcanic and metavolcaniclastic rocks that crop out in the northeastern Maryland Piedmont and was correlated with layered rocks of the Wilmington Complex. Ward's (1959)

Wilmington Complex amphibolite units were placed in the James Run Formation by Pickett (1976), Hager (1976), and Thompson (1979). Higgins (1990) traced the Gilpins Falls and Big Elk Members of the James Run Formation into Delaware, where the belt of volcanic rocks is truncated by the pluton composed of Christianstead Gneiss at the southwest end of the Wilmington Complex (Fig. 2). Schenck et al. (2000) assigned the Christianstead, Windy Hills, and Faulkland Gneisses to the Wilmington Complex. However, as presented here, geochronologic data preclude correlation of metavolcanic rocks of the James Run Formation in Maryland with metavolcanic units now included within the Wilmington Complex.

DESCRIPTION OF DATED UNITS WITHIN THE WILMINGTON COMPLEX

The units that comprise the Wilmington Complex are described in detail in the report (Plank et al., 2000) accompanying the bedrock geologic map of Delaware and adjacent Pennsylvania (Schenck et al., 2000). Below is a brief description of the units from which samples were obtained for U-Pb geochronology.

The Brandywine Blue Gneiss (Plank et al., 2000) is the name applied to the coarse-grained, granulite-grade, felsic units within the Wilmington Complex (Fig. 2). This name is used because freshly quarried rock along Brandywine Creek is bright blue to bluish-gray in color. On weathered surfaces, the rock is light gray. In general, the unit is a homogeneous, well-lineated, light-gray, felsic gneiss containing discontinuous layers of mafic rocks. It is composed primarily of quartz and plagioclase, \pm orthopyroxene, clinopyroxene, and brown-green amphibole (Plank et al., 2000). Major and trace element compositions of the mafic layers suggest an island-arc origin (Plank et al., 2001). The fairly homogeneous character of the rock implies that the protolith was an intrusive rock, although no igneous textures or structures have been preserved. A volcanic or subvolcanic origin also is possible. Petrologic studies (Wagner and Srogi, 1987; Srogi, 1988; Srogi et al., 1993) indicate that the Brandywine Blue Gneiss attained granulite-grade metamorphism with temperatures and pressures of 800 ± 50 °C and 5–6 kbar, resulting in limited anatexis as shown by local veins and leucosomes.

The Rockford Park Gneiss is composed of well-foliated, interlayered mafic and felsic gneisses. Overall, it is more mafic and exhibits more layering than the Brandywine Blue Gneiss. Together, both units comprise the granulite-grade "banded gneiss" of Ward (1959). The

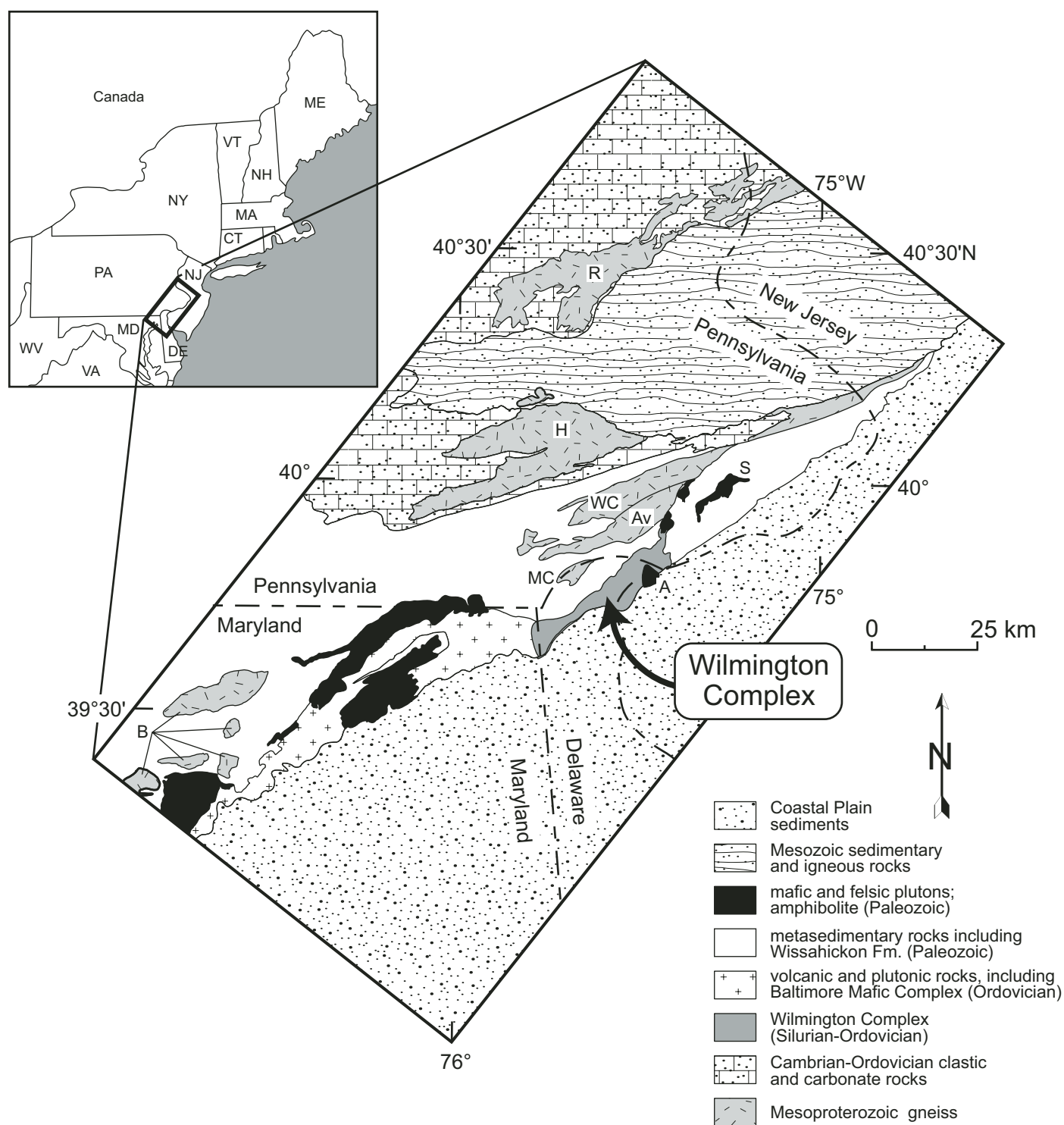


Figure 1. Simplified regional geologic map showing location of the Wilmington Complex. Modified from Berg et al. (1980), Hanan and Sinha (1989), and Fail (1997). Abbreviations: A—Arden Plutonic Supersuite, Av—Avondale massif, B—Baltimore Gneiss domes, H—Honeybrook Upland, MC—Mill Creek Nappe, R—Reading Prong, WC—West Chester massif, S—Springfield pluton.

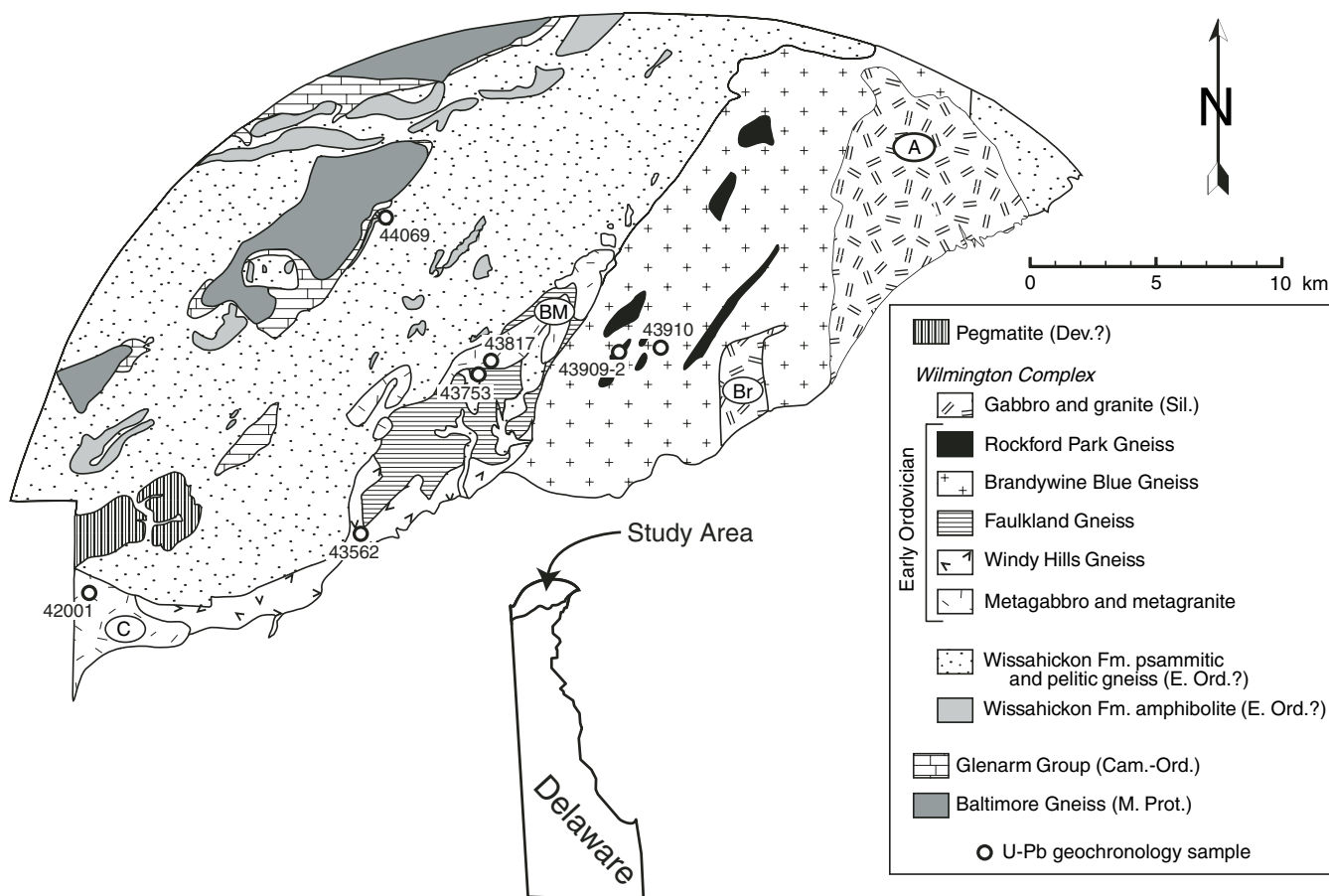


Figure 2. Geologic map of crystalline rocks of northern Delaware Piedmont (modified from Schenck et al., 2000). A—Arden Plutonic Supersuite, BM—Barley Mill Gneiss, Br—Bringinghurst Gabbro, C—Christianstead Gneiss. Locations of geochronology samples are shown by circles.

mafic gneisses, composed of plagioclase, orthopyroxene, clinopyroxene, and hornblende, plus minor quartz and biotite (Plank et al., 2000), have chemical signatures suggestive of boninitic affinity, implying, by analogy with modern tectonic settings (Kim and Jacobi, 2002), a forearc origin (Plank et al., 2001). The felsic layers are composed of quartz and plagioclase, plus minor orthopyroxene \pm clinopyroxene (Plank et al., 2000). Based on two-pyroxene geothermometry, metamorphic temperatures of $\sim 800 \pm 50$ °C were attained (Wagner and Srogi, 1987; Srogi, 1988), resulting in local anatexis. Close spatial association of the Rockford Park Gneiss and the enveloping Brandywine Blue Gneiss (Fig. 2) suggests that the two units were metamorphosed together to granulite grade.

The Windy Hills Gneiss is composed of interlayered amphibolite, felsic gneiss, and biotite-quartz-plagioclase gneiss of inferred volcanic and volcanoclastic origin (Fig. 2). It is interpreted as being interlayered with metasedimentary rocks mapped as Wissahickon Formation (Plank et al., 2000), although no contacts

have been observed. The sample of Windy Hills Gneiss collected for U-Pb geochronology is a cordierite-gedrite gneiss (apparently highly altered from its original felsic volcanic protolith). Rocks of the Windy Hills Gneiss have been metamorphosed to amphibolite facies. Subtle mineralogic changes suggest increasing grade of metamorphism northeastward approaching the Brandywine Blue Gneiss. Although similar in appearance and mineralogy to mafic rocks of the Rockford Park Gneiss, the compositions of mafic layers within the Windy Hills Gneiss suggest an island-arc tholeiitic origin (Plank et al., 2001).

The Faulkland Gneiss primarily is an amphibolitic unit, with minor felsic layers, and is interlayered with metasedimentary rocks mapped as the Wissahickon Formation (Fig. 2). The amphibolites are composed of hornblende and plagioclase, plus minor quartz, orthopyroxene, and clinopyroxene (Plank et al., 2000). Chemical compositions suggest that the protolith of the amphibolite, interpreted as volcanic and volcanoclastic in origin, had an arc affinity (Plank et

al., 2001). The Faulkland Gneiss was metamorphosed to upper amphibolite facies conditions.

The Christianstead Gneiss is a moderately deformed, coarse-grained granodioritic gneiss. It is composed of quartz, plagioclase, microcline, and biotite (Plank et al., 2000). It intrudes the Windy Hills Gneiss and is interpreted to intrude the Wissahickon Formation. However, its field relationship to the Wissahickon is unknown, because the contact is not exposed. It is metamorphosed to amphibolite facies conditions.

The Barley Mill Gneiss occurs as a composite pluton, composed of tonalitic, trondhjemitic, and granodioritic gneisses. The dated sample of tonalitic gneiss is composed of plagioclase, quartz, hornblende, and biotite (Plank et al., 2000). It intrudes the Faulkland Gneiss and Wissahickon Formation and is interpreted to intrude the Brandywine Blue Gneiss. Although it is adjacent to the granulite-grade Brandywine Blue Gneiss, it was only metamorphosed to amphibolite facies conditions.

The Arden Plutonic Supersuite includes gabbro, diorite, quartz diorite, quartz monzonite,

granodiorite, and granite (Srogi, 1988; Srogi and Lutz, 1997). On the bases of occurrence, mineralogy, and chemical composition, the Arden Plutonic Supersuite has been subdivided into the Ardentown Granitic Suite, Perkins Run Gabbro-norite Suite, and Biotite Tonalite (Plank et al., 2000). These igneous rocks, which retain igneous textures, but are moderately foliated, intrude the Brandywine Blue Gneiss and Wissahickon Formation (Fig. 2). Fabrics in the Arden Plutonic Supersuite may be flow structures related to emplacement, or due to syn- or post-crystallization regional deformation. The dated sample, a pyroxene quartz diorite, is a member of the Ardentown Granitic Suite. It is composed of quartz, K-feldspar, orthopyroxene, clinopyroxene, and biotite.

The Wissahickon Formation consists of a sequence of complexly folded metasedimentary rocks, primarily psammitic and pelitic gneisses and schists, plus amphibolite, which occur west, northwest, and east of the Wilmington Complex (Fig. 2). Different rock types, presumably from dissimilar tectonic settings, have been grouped within the Wissahickon. For example, Wagner and Srogi (1987) hypothesized that the Wissahickon is composed of sediments from an accretionary prism, forearc, and collisional flysch. Two types of amphibolite occur within the Wissahickon Formation, suggesting protolith origins as mid-ocean-ridge basalt (MORB) and intraplate basalt (Plank et al., 2001). Blackmer and Bosbyshell (2002) suggested that the amalgamation of a diverse suite of schists into the Wissahickon Formation is an oversimplification of the stratigraphy.

The relationship between the Wilmington Complex and nearby rocks mapped as Wissahickon Formation is controversial. Rocks of the Wissahickon Formation in Delaware have been metamorphosed to upper amphibolite facies conditions, with metamorphic grade increasing southeast toward the Brandywine Blue Gneiss. Metavolcanic rocks of the Faulkland and Windy Hills Gneisses are interlayered with metasedimentary rocks and amphibolites assigned to the Wissahickon Formation. Plank et al. (2000) suggested that the protolith of the Brandywine Blue Gneiss intruded into the Wissahickon Formation. However, the contact between the Brandywine Blue Gneiss and the Wissahickon Formation is poorly exposed, and the nature of that contact in Delaware is unclear. Plutonic rocks of the Brandywine Blue Gneiss and the Barley Mill Gneiss contain xenoliths of pelitic gneiss that may or may not have been derived from the Wissahickon. Additional field evidence in support of an intrusive relationship occurs in Pennsylvania, north of the map area in Figure 2, where a mafic dike essentially identical

in composition to mafic Rockford Park Gneiss crosscuts an outcrop of both the Wissahickon Formation and Ward's (1959) "banded gneiss" of the Wilmington Complex (Bosbyshell et al., 1999). This dike intruded both units before metamorphism and deformation occurred at 440–430 Ma (Bosbyshell et al., 1999, 2000).

U-Pb GEOCHRONOLOGY

Previous Age Data

Because rocks of the Wilmington Complex have been metamorphosed at upper amphibolite to granulite facies, the determination of protolith ages for the gneisses has been very difficult. Pioneering work by Grauert and Wagner (1974, 1975) recognized the complexities of zircons from rocks now called the Brandywine Blue Gneiss (the "banded gneiss" of Ward, 1959). Conventional U-Pb analyses of multigrain populations of zircon yielded a linear array of data with lower and upper concordia intercept ages of ca. 441 and 1500 Ma (recalculated using Isoplot/Excel [Ludwig, 2003] to 441 ± 5 and 1522 ± 270 Ma). All of the isotopic data plot near the lower end of the discordia, including one fraction that is concordant at ca. 441 Ma. Using cathodoluminescence (CL), Grauert and Wagner (1975) showed that most of the zircons were composed of distinct cores and overgrowths, interpreted as igneous and metamorphic, respectively, in origin. They concluded that the lower intercept age of 441 Ma was the time at which the gneisses were metamorphosed to granulite facies. No estimates of the protolith age or interpretation of the upper intercept age were made.

Foland and Muessig (1978) analyzed several samples from the Arden pluton (now called the Ardentown Granitic Suite of the Arden Plutonic Supersuite) for Rb and Sr isotopes, deriving a whole-rock isochron age of 502 ± 20 Ma (2σ) (recalculated to 519 ± 27 Ma). Rb-Sr data from several whole rock-biotite pairs yielded ages of ca. 368 Ma. Bosbyshell et al. (1998) reported a slightly discordant ion microprobe U-Pb age of 422 ± 7 Ma for zircon from quartz norite of the Ardentown Granitic Suite. Foland (1979, 1983) dated several samples of biotite from the Arden pluton by both K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ isotopic techniques. Results ranged from 363 ± 5 to 572 ± 8 Ma (1σ); three analyses yielded plateau ages of ca. 435 Ma. These data were interpreted to indicate that the time of closure for Ar diffusion in biotite from the Arden pluton was at 363 ± 5 Ma (Foland, 1979, 1983), in agreement with the Rb-Sr whole rock-biotite data. All older ages, including biotite from the "banded gneiss" dated at 431 ± 6 Ma, were thought to be due to

excess ^{40}Ar and therefore were considered to be geologically meaningless.

On the bases of the presence of igneous textures, only slight deformational fabrics, and intrusive relations, rocks of the Arden Plutonic Supersuite are considered to be younger than the strongly deformed, granulite facies Brandywine Blue Gneiss of the Wilmington Complex. However, as pointed out by Wagner and Srogi (1987) and Srogi et al. (1993), the ages of the Arden and Wilmington Complex, based on Rb-Sr whole rock data from Foland and Muessig (1978) and U-Pb zircon data from Grauert and Wagner (1975), respectively, were problematic, because the granulite facies metamorphism of the Wilmington Complex supposedly occurred at ca. 441 Ma (Grauert and Wagner, 1975), which is inconsistent with an age of 502 ± 20 Ma for the relatively undeformed Arden Plutonic Supersuite. As will be shown below, our new SHRIMP U-Pb zircon ages for both the Arden Plutonic Supersuite and the Brandywine Blue Gneiss are younger than the 502 ± 20 Ma Rb-Sr age of Foland and Muessig (1978). Perhaps coincidentally, the Silurian ages of some of the biotite samples from both rocks are now in agreement with U-Pb zircon data for the timing of emplacement of the Arden Plutonic Supersuite and metamorphism of the Wilmington Complex.

Methods

Zircon and monazite were extracted from rock samples of ~10–15 kg using a crusher, pulverizer, Wilfley table, magnetic separator, and methylene iodide. Individual grains were hand-picked under a binocular microscope, mounted in epoxy, hand-ground to about half-thickness with 1500 grit wet-dry sand paper, and polished on a lap wheel with 6 μm and 1 μm diamond suspension. Prior to analysis, all grains were photographed in transmitted and reflected light. In addition, zircon was imaged by CL, whereas monazite was imaged by backscattered electrons (BSE) on the scanning electron microscope, in order to identify pristine areas for analysis, and to determine if multiple age components (i.e., cores and overgrowths) were present.

Zircon and monazite were dated using the SHRIMP II and SHRIMP-RG at the Research School of Earth Sciences (RSES), Australian National University (ANU), and the SHRIMP-RG at the U.S. Geological Survey (USGS)-Stanford isotope facility (Table 1). Analytical procedures for zircon dating followed methods described in Williams (1998). SHRIMP II analyses of monazite sample 43753 (session 1) (following the methodology of Williams et al., 1996) utilized energy filtering to remove lower-energy molecules that cause

TABLE 1. SHRIMP U-Th-Pb DATA FOR ZIRCON AND MONAZITE FROM ROCKS OF THE WILMINGTON COMPLEX AND WISSAHICKON FORMATION, AS MAPPED IN DELAWARE

Sample [†] (location)	Measured $\frac{^{204}\text{Pb}}{^{206}\text{Pb}}$	Measured $\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	Common $\frac{^{206}\text{Pb}}{^{206}\text{Pb}}$ (%)	U (ppm)	Th/U	$\frac{^{206}\text{Pb}^{\dagger}}{^{238}\text{U}}$ (Ma)	Error [§] (Ma)	$\frac{^{238}\text{U}^{\dagger}}{^{206}\text{Pb}}$	Error [§] (%)	$\frac{^{207}\text{Pb}^{\dagger}}{^{206}\text{Pb}}$	Error [§] (%)	$\frac{^{207}\text{Pb}^{\dagger, \ddagger}}{^{206}\text{Pb}}$ (Ma)	Error [§] (Ma)
A. 43910 (Brandywine Blue Gneiss)													
910-1.1c	0.000021	0.0532	—	220	0.27	452.1	5.4	13.82	1.2	0.0529	1.8		
910-2.1c	—	0.0569	0.13	91	0.26	443.4	6.1	14.02	1.4	0.0575	2.6		
910-3.1c	—	0.0558	0.04	259	0.68	434.7	5.2	14.33	1.2	0.0562	2.1		
910-4.1c	0.0000078	0.0575	0.14	165	0.33	468.1	5.8	13.26	1.3	0.0574	2.5		
910-4.2r	0.000017	0.0562	—	2916	0.16	483.1	5.2	12.86	1.1	0.0559	0.4		
910-5.1c	0.00013	0.0568	0.08	114	0.23	459.2	6.1	13.57	1.4	0.0549	3.7		
910-6.1c	0.000045	0.0556	0.05	99	0.49	418.6	5.7	14.91	1.4	0.0549	2.6		
910-6.2r	0.000013	0.0557	0.04	244	0.42	427.1	5.1	14.6	1.2	0.0555	2		
910-7.1c	—	0.0584	0.23	145	0.22	475	6.1	13.03	1.3	0.0596	1.9		
910-8.1r	0.00045	0.0596	0.5	46	0.53	434.7	7.2	14.38	1.8	0.0531	11.6		
910-8.2c	0.0012	0.0722	1.93	127	0.24	474.5	7.4	13.13	1.8	0.054	13.7		
910-9.1c	0.00012	0.0561	0.09	194	0.72	426.6	6.1	14.63	1.5	0.0543	2.7		
910-9.2r	0.00084	0.0632	0.94	40	0.5	435.6	7.3	14.39	1.8	0.0509	11.8		
910-10.1c	0.00012	0.0574	0.14	114	0.26	466	6.3	13.35	1.4	0.0556	3		
910-11.1c	0.00045	0.0613	0.79	57	0.41	410	6.6	15.23	1.7	0.0548	7.1		
910-12.1c	0.00072	0.0587	0.43	65	0.37	421.5	6.5	14.93	1.7	0.0481	11.2		
910-13.1c	0.00051	0.0563	0.1	73	0.53	431.9	6.4	14.55	1.6	0.0488	7.8		
910-14.1c	0.00012	0.0556	0.02	179	0.57	432.1	7.1	14.45	1.7	0.0539	2.9		
910-15.1eqc	0.00016	0.056	—	85	0.22	480.4	6.8	12.97	1.4	0.0536	3.7		
910-16.1eqc	0.000072	0.057	0.06	182	0.24	472.9	5.9	13.15	1.3	0.0559	2.4		
910-17.1eqc	0.00017	0.0573	0.07	216	0.2	480.5	5.9	12.95	1.2	0.0548	2.8		
910-18.1eqc	0.000082	0.0585	0.18	247	0.25	494.2	6	12.55	1.2	0.0573	1.6		
910-19.1eqc	—	0.059	0.27	80	0.21	485.4	7	12.74	1.5	0.0598	2.9		
910-20.1eqc	0.000021	0.0576	0.13	98	0.23	473.4	6.5	13.11	1.4	0.0573	3.4		
910-21.1eqc	0.0001	0.0585	0.2	206	0.22	486.7	6	12.75	1.3	0.057	2.2		
910-22.1eqc	0.00022	0.0564	0.04	127	0.26	453	7.2	13.78	1.6	0.0532	4		
910-23.1eqc	0.0001	0.0574	0.2	131	0.24	443.5	5.8	14.04	1.3	0.0559	2.7		
910-24.1eqc	0.00004	0.0566	0.04	177	0.35	466.8	5.8	13.32	1.3	0.0561	2.4		
910-25.1c	0.00031	0.0568	0.2	182	0.63	419.5	5.4	14.93	1.3	0.0523	2.6		
910-26.1c	—	0.059	0.51	120	0.56	407.9	5.7	15.22	1.4	0.0599	2.5		
910-27.1c	—	0.0556	0.03	104	0.48	427.5	7	14.58	1.7	0.0556	2.5		
910-28.1c	0.00045	0.0557	0.02	120	0.48	434	5.9	14.47	1.4	0.049	6.7		
910-29.1c	0.00011	0.0576	0.28	105	0.48	424.5	5.9	14.68	1.4	0.056	3.1		
B. 43909-2 (Rockford Park Gneiss)													
909-1.1c	0.000059	0.0571	0.22	198	0.62	425.4	5.3	14.64	1.3	0.0562	2.5		
909-2.1c	0.000056	0.0564	—	272	1	479.5	5.9	12.97	1.2	0.0556	2		
909-3.1c	0.000052	0.0577	0.2	258	0.87	456.5	6.9	13.61	1.5	0.057	2		
909-4.1c	0.000068	0.0569	0.04	256	0.59	474.2	5.9	13.11	1.3	0.0559	1.9		
909-4.2r	0.00014	0.0583	0.37	202	0.58	426.5	6.3	14.6	1.5	0.0563	2.2		
909-5.1c	0.0000068	0.0561	0.07	318	0.2	432.5	5.1	14.4	1.2	0.056	1.8		
909-6.1c	0.00014	0.0572	0.07	245	0.84	476.6	5.8	13.06	1.2	0.0551	1.9		
909-7.1c	0.00019	0.0572	0.08	254	0.65	473.1	6.2	13.17	1.3	0.0543	2.3		
909-8.1c	0.00013	0.0559	—	196	0.89	477	6.5	13.06	1.4	0.054	2		
909-9.1c	0.000095	0.0556	—	184	0.51	506.1	7	12.29	1.4	0.0542	2.3		
909-10.1c	0.000099	0.0572	0.06	262	1.04	478.5	5.9	12.99	1.3	0.0557	1.6		
909-11.1c	0.000002	0.0573	0.13	169	0.62	464.5	5.9	13.37	1.3	0.0573	2.1		
909-12.1c	0.0000076	0.0588	0.27	177	0.49	476.9	6	12.99	1.3	0.0587	2.2		
909-13.1c	0.000038	0.0577	0.17	304	0.9	466.6	5.5	13.31	1.2	0.0571	1.7		
909-14.1c	0.00018	0.0608	0.59	272	0.84	451.4	5.4	13.75	1.2	0.0581	2.4		
909-15.1c	—	0.0561	—	226	0.6	485.9	5.9	12.78	1.2	0.0565	1.8		
909-16.1c	0.00004	0.0581	0.2	169	0.81	472.6	6.2	13.13	1.3	0.0576	2.2		
909-16.2r	0.00016	0.0565	—	176	0.47	480.1	6.2	12.97	1.3	0.0543	3.2		
909-17.1c	—	0.0568	—	163	0.77	506.8	11.7	12.23	2.4	0.057	1.9		
909-18.1c	0.0001	0.0567	—	213	1.11	480.7	6.7	12.94	1.4	0.0552	2.4		
909-19.1eqc	0.00022	0.0562	0.06	154	0.66	442.1	5.7	14.14	1.3	0.0529	2.6		
909-20.1eqc	0.000074	0.0557	0.01	352	0.62	435.9	5.1	14.31	1.2	0.0546	1.5		
909-21.1eqc	0.000054	0.0563	0.14	330	0.85	421.4	5	14.8	1.2	0.0556	2		
909-22.1eqc	0.00011	0.0556	—	197	0.74	441.3	5.6	14.14	1.3	0.054	2.1		
909-23.1eqc	0.000059	0.0564	0.11	569	0.91	434.8	5	14.33	1.2	0.0556	1.2		
909-24.1eqc	0.000042	0.0582	0.34	243	0.83	430.9	5.3	14.43	1.2	0.0576	1.8		

(continued)

SHRIMP U-Pb GEOCHRONOLOGY OF THE WILMINGTON COMPLEX, DELAWARE

TABLE 1. SHRIMP U-Th-Pb DATA FOR ZIRCON AND MONAZITE FROM ROCKS OF THE WILMINGTON COMPLEX AND WISSAHICKON FORMATION, AS MAPPED IN DELAWARE (*continued*)

Sample [†] (location)	Measured $\frac{^{204}\text{Pb}}{^{206}\text{Pb}}$	Measured $\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	Common $\frac{^{206}\text{Pb}}{^{206}\text{Pb}}$ (%)	U (ppm)	Th/U	$\frac{^{206}\text{Pb}^{\dagger}}{^{238}\text{U}}$ (Ma)	Error [§] (Ma)	$\frac{^{238}\text{U}^{\#}}{^{206}\text{Pb}}$	Error [§] (%)	$\frac{^{207}\text{Pb}^{\#}}{^{206}\text{Pb}}$	Error [§] (%)	$\frac{^{207}\text{Pb}^{\dagger, \ddagger}}{^{206}\text{Pb}}$ (Ma)	Error [§] (Ma)
C. 43562 (Windy Hills Gneiss)													
562-1.1c	0.000073	0.0575	0.08	144	0.82	485.5	6.3	12.79	1.3	0.0565	2.5		
562-2.1c	0.00013	0.0591	0.31	219	0.6	474.6	5.8	13.08	1.2	0.0571	2.4		
562-3.1c	0.0000017	0.0578	0.12	138	0.36	486.1	6.3	12.75	1.3	0.0578	2.2		
562-4.1c	0.00071	0.056	—	73	2.01	486.8	7.4	12.93	1.6	0.0456	9.2	1238	15
562-5.1ctz	0.000024	0.082	—	293	0.34	1249	14.1	4.68	1.2	0.0817	0.8		
562-6.1c	0.000006	0.0605	0.47	159	0.58	480.2	6.1	12.87	1.3	0.0604	1.8		
562-7.1ctz	0.00012	0.0683	—	120	1.32	925.5	11.6	6.5	1.3	0.0666	1.6		
562-8.1c	0.00019	0.1138	0.08	133	1.16	1850.5	27.1	3.01	1.5	0.1113	2	1821	36
562-9.1c	0.0012	0.0606	0.27	100	0.66	544.6	8.4	11.56	2.2	0.0427	30.7		
562-10.1c	—	0.0579	0.17	130	0.89	474.3	7.7	13.07	1.7	0.0584	2.1		
562-11.1c	0.00031	0.0603	0.48	73	0.84	468.1	8	13.29	1.8	0.0558	5.6		
562-12.1c	0.000039	0.0563	—	814	0.18	485.5	5.4	12.8	1.1	0.0558	0.9		
562-13.1ctz	—	0.0762	0.29	127	1.11	1035.4	12.8	5.72	1.3	0.0762	1.2	1100	25
562-14.1c	0.0004	0.0599	0.42	68	0.49	471.9	7	13.21	1.6	0.0541	6.4		
562-15.1ctz	0.000068	0.0795	—	176	0.9	1211.8	15.7	4.85	1.3	0.0785	1.1	1160	21
562-16.1c	0.00007	0.0593	0.27	164	0.89	493.5	6.3	12.55	1.3	0.0582	2.4		
562-17.1ctz	0.000024	0.0816	0.12	301	0.49	1212.3	13.7	4.83	1.2	0.0813	0.7	1229	14
562-18.1ctz	0.00014	0.0745	0.19	128	0.87	1010.4	12.5	5.9	1.3	0.0725	1.6	1000	33
562-19.1ctz	—	0.0588	—	259	0.14	568.1	7.2	10.85	1.3	0.0595	1.5		
562-19.2r	0.000027	0.0573	0.05	214	0.08	485.3	6.7	12.79	1.4	0.0569	2.4	1206	31
562-20.1ctz	—	0.0802	0.38	140	0.32	1119.3	16.1	5.25	1.5	0.0804	1.6		
562-21.1c	0.00015	0.0584	0.2	130	1.02	484.8	6.4	12.81	1.3	0.0562	2.8	1018	22
562-22.1eq-r	0.000073	0.0742	0.06	212	0.74	1033.1	12.1	5.76	1.2	0.0731	1.1	1001	22
562-23.1eq-ctz	0.000027	0.0729	—	231	0.28	1052.5	17.3	5.65	1.7	0.0726	1.1		
562-24.1c	0.00014	0.0591	0.31	144	0.34	474.2	7.4	13.09	1.6	0.0571	2.3		
562-25.1eq-ctz	0.00004	0.0815	—	246	0.81	1261.3	14.5	4.64	1.2	0.0809	0.9	1220	17
562-25.2eq-r	0.000012	0.0813	0.04	636	0.8	1221	13.4	4.79	1.1	0.0811	0.7	1225	14
D. 43753 (Faulkland Gneiss)													
753-1.1c	0.000043	0.0579	0.14	168	0.66	481.2	6.2	12.89	1.3	0.0572	3.2		
753-2.1c	0.00015	0.0572	0.12	167	0.55	462.4	6.3	13.47	1.4	0.055	2.8		
753-3.1c	0.000011	0.0568	—	183	0.77	486.4	6.7	12.76	1.4	0.0566	2.6		
753-4.1c	—	0.057	0.05	633	2.2	476.5	5.4	13.03	1.2	0.0571	1		
753-5.1c	0.0001	0.0614	0.55	116	1.13	488.7	6.6	12.65	1.4	0.06	2.8		
753-6.1c	—	0.0587	0.21	480	0.55	491.4	5.7	12.6	1.2	0.0589	1.3		
753-7.1c	0.000029	0.0573	0.07	265	0.54	480.9	5.8	12.91	1.2	0.0569	1.7		
753-8.1c	0.000029	0.0579	0.21	178	0.75	458.8	5.8	13.54	1.3	0.0574	2.5		
753-9.1c	0.00025	0.0799	0.12	86	0.38	1169.4	17.5	5.04	1.5	0.0764	2.1	1106	41
753-10.1c	0.00081	0.07	1.96	615	0.69	377.6	38.8	16.49	10.5	0.0582	6.2		
753-11.1c	0.00013	0.0569	0.02	264	0.6	478.8	6.2	13	1.3	0.0549	2.6		
753-12.1c	0.000087	0.0576	0.09	467	0.56	489.3	5.6	12.69	1.2	0.0564	1.4		
753-13.1c	0.00022	0.0585	0.26	208	0.53	468.2	5.8	13.29	1.3	0.0553	2.6		
753-14.1c	—	0.0559	—	229	0.06	454.9	18.6	13.68	4.2	0.056	1.7		
753-15.1c	0.000072	0.0727	0.94	240	0.16	772.5	9.1	7.79	1.2	0.0717	1.2		
753-16.1c	0.00015	0.0574	0.09	278	0.63	480.8	5.8	12.94	1.2	0.0553	1.6		
753-17.1eq-c	0.000037	0.0579	0.14	128	0.88	481.3	7.9	12.89	1.7	0.0573	2.8		
753-18.1eq-c	—	0.058	0.15	174	0.86	484.9	6.3	12.78	1.3	0.0581	1.8		
753-19.1eq-c	0.000022	0.0581	0.19	163	0.7	474.9	6.1	13.06	1.3	0.0578	2.2		
(Monazite, session 1)													
753m-1.1r-uz	—	0.0566	0.23	1101	33.8	405.1	7.5	15.38	1.9	0.0566	1.5		
753m-1.2c-uz	0.000030	0.0556	0.09	5715	0.7	410.4	6.8	15.21	1.7	0.0552	0.7		
753m-2.1r-uz	0.000041	0.0560	0.06	4332	4.9	432.3	7.4	14.42	1.8	0.0554	0.8		
753m-2.2c-sz	0.000000	0.0554	0.00	5143	0.8	429.8	6.6	14.50	1.6	0.0554	0.9		
753m-3.1c-oz	0.000046	0.0551	—	5011	10.8	434.5	6.8	14.36	1.6	0.0545	0.9		
753m-3.2r-uz	—	0.0556	—	4439	7.1	490.1	7.8	12.67	1.6	0.0560	1.0		
753m-4.1r-uz	0.000030	0.0561	0.10	2754	1.9	422.2	6.8	14.77	1.6	0.0556	0.9		
753m-4.2r-uz	0.000047	0.0552	—	3366	10.4	446.0	7.1	13.98	1.6	0.0545	1.0		
753m-4.3c-oz	0.000018	0.0553	0.01	2942	10.7	419.1	6.8	14.89	1.6	0.0550	1.0		
753m-4.4r-uz	0.000000	0.0561	0.03	3105	11.0	449.3	7.3	13.85	1.6	0.0561	0.9		
753m-5.1i-oz	0.000016	0.0562	0.12	3530	6.2	421.3	6.7	14.79	1.6	0.0560	0.8		

(continued)

TABLE 1. SHRIMP U-Th-Pb DATA FOR ZIRCON AND MONAZITE FROM ROCKS OF THE WILMINGTON COMPLEX AND WISSAHICKON FORMATION, AS MAPPED IN DELAWARE (*continued*)

Sample [†] (location)	Measured $\frac{^{204}\text{Pb}}{^{206}\text{Pb}}$	Measured $\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	Common $\frac{^{206}\text{Pb}}{^{206}\text{Pb}}$ (%)	U (ppm)	Th/U	$\frac{^{206}\text{Pb}^{\dagger}}{^{238}\text{U}}$ (Ma)	Error [§] (Ma)	$\frac{^{238}\text{U}^{\dagger}}{^{206}\text{Pb}}$	Error [§] (%)	$\frac{^{207}\text{Pb}^{\dagger}}{^{206}\text{Pb}}$	Error [§] (%)	$\frac{^{207}\text{Pb}^{\dagger, \ddagger}}{^{206}\text{Pb}}$ (Ma)	Error [§] (Ma)
753m-6.1c-sz	0.000073	0.0573	0.35	1222	19.6	392.3	7.1	15.90	1.8	0.0563	1.7		
753m-7.1c-oz	0.000029	0.0554	—	3378	7.1	444.5	7.1	14.02	1.6	0.0550	0.9		
753m-8.1c-oz	0.000018	0.0564	0.17	3103	1.5	413.3	7.1	15.08	1.8	0.0561	0.9		
753m-9.1c-uz	0.000003	0.0557	0.09	2563	11.5	411.7	6.7	15.15	1.7	0.0557	1.1		
753m-9.2r-uz	0.000000	0.0556	0.03	3820	8.4	428.1	6.8	14.56	1.6	0.0556	1.0		
753m-10.1r-uz	0.000121	0.0552	0.01	1326	25.4	417.2	7.4	14.99	1.8	0.0535	2.4		
753m-10.2c-uz	0.000026	0.0550	—	2399	13.1	433.7	7.1	14.39	1.7	0.0547	1.1		
753m-11.1c-oz	0.000015	0.0558	0.06	5312	1.8	424.6	7.0	14.69	1.7	0.0555	0.7		
753m-12.1c-uz	0.000039	0.0544	—	4093	1.4	441.7	6.9	14.13	1.6	0.0539	0.8		
753m-12.2r-sz	0.000076	0.0558	0.09	2735	6.0	414.8	6.7	15.05	1.6	0.0547	1.2		
753m-13.1r-uz	0.000017	0.0554	—	3357	8.5	431.1	6.9	14.47	1.6	0.0551	0.9		
753m-13.2c-sz	0.000030	0.0553	—	3378	6.4	433.3	6.9	14.39	1.6	0.0549	0.9		
753m-14.1r-uz	0.000029	0.0555	0.08	2041	14.0	408.2	7.7	15.29	1.9	0.0551	1.2		
753m-14.2c-uz	0.000033	0.0561	0.08	2754	11.6	431.7	7.4	14.44	1.7	0.0556	1.1		
753m-15.1c-sz	0.000125	0.0558	0.01	1352	24.6	440.8	7.8	14.16	1.8	0.0540	1.6		
753m-15.2r-uz	0.000029	0.0564	0.12	5107	2.1	428.7	6.6	14.53	1.6	0.0560	0.7		
753m-16.1c-sz	0.000061	0.0560	0.14	2437	10.1	407.6	6.7	15.31	1.7	0.0551	1.2		
(Monazite, session 2)													
753m-1.1c-uz	0.000334	0.0570	0.20	1930	20.8	426.4	2.9	14.68	0.7	0.0521	2.5		
753m-2.1c-uz	0.000140	0.0560	0.07	4088	11.7	429.0	2.3	14.56	0.5	0.0540	1.1		
753m-3.1i-uz	0.000168	0.0563	0.15	2686	24.9	414.5	2.6	15.08	0.6	0.0538	1.5		
753m-3.2c-uz	0.000196	0.0553	—	3791	15.3	447.8	2.5	13.96	0.6	0.0525	1.7		
753m-3.3i-uz	0.000160	0.0554	0.07	3692	13.1	408.6	2.2	15.32	0.5	0.0531	1.2		
753m-4.1c-uz	0.000042	0.0558	0.05	5913	0.3	431.1	2.0	14.46	0.5	0.0552	0.6		
753m-5.1c-sz	0.000130	0.0552	0.06	5101	9.3	400.1	2.0	15.64	0.5	0.0533	1.0		
753m-6.1c-uz	0.000093	0.0557	0.06	4987	8.3	421.6	4.7	14.81	1.1	0.0543	0.9		
753m-7.1c-sz	0.000220	0.0569	0.25	2611	19.0	409.0	4.9	15.29	1.2	0.0537	1.6		
753m-8.1c-uz	0.000196	0.0556	0.11	3214	17.2	399.1	4.7	15.69	1.2	0.0527	1.4		
753m-8.2r-sz	0.000102	0.0554	0.13	6852	10.5	387.6	4.3	16.14	1.1	0.0539	0.8		
753m-9.1c-sz	0.000149	0.0558	0.13	4864	9.7	402.9	4.6	15.53	1.2	0.0536	1.2		
753m-10.1c-uz	0.000040	0.0557	0.04	5863	5.3	428.0	4.8	14.57	1.1	0.0551	0.7		
753m-11.1c-oz	0.000143	0.0549	—	5959	7.8	449.3	5.0	13.91	1.1	0.0528	1.0		
E. 42001 (Christianstead Gneiss)													
001-1.1c	—	0.0572	0.05	260	0.5	484.1	9	12.82	1.9	0.0572	2		
001-2.1c	—	0.0562	—	204	0.42	487.9	10.4	12.73	2.2	0.0562	2		
001-3.1c	0.000068	0.0551	—	226	0.46	510.6	9.2	12.18	1.9	0.0541	1.9		
001-4.1c	0.00014	0.057	0.03	164	0.51	481.2	9.1	12.93	1.9	0.0549	3.7		
001-5.1c	—	0.055	—	207	0.54	481.6	9	12.92	1.9	0.055	2.3		
001-6.1c	—	0.054	—	250	0.58	479.7	8.8	12.99	1.9	0.054	1.9		
001-7.1c	—	0.0567	—	523	0.39	484.3	9.3	12.82	2	0.0567	1.3		
001-8.1c	—	0.0568	—	484	0.49	491	8.7	12.64	1.8	0.0568	1.3		
001-9.1c	—	0.0563	—	258	0.44	479.6	8.7	12.95	1.9	0.0563	1.7		
001-10.1c	—	0.0573	0.03	516	0.47	494.9	9.5	12.53	2	0.0573	1.1		
001-11.1c	—	0.0568	0.02	335	0.43	476	8.7	13.05	1.9	0.0568	1.5		
001-12.1c	0.000037	0.0558	—	441	0.57	485.9	8.6	12.8	1.8	0.0553	1.4		
001-13.1c	—	0.0564	—	381	0.41	496.1	9.3	12.51	1.9	0.0564	1.5		
001-14.1c	—	0.0569	0.11	334	0.38	453.1	17.5	13.72	4	0.0569	1.5		
001-15.1c	—	0.0575	0.1	229	0.58	476.1	8.7	13.03	1.9	0.0575	1.8		
001-16.1c	—	0.0568	0.01	455	0.45	478	9.3	12.99	2	0.0568	1.3		
001-17.1c	0.000034	0.0572	0.05	305	0.75	484.1	9	12.83	1.9	0.0567	1.5		
001-18.1c	—	0.057	0.03	224	0.78	482	8.7	12.88	1.8	0.057	1.7		
F. 43817 (Barley Mill Gneiss)													
817-1.1c	—	0.0563	—	272	0.58	482.2	8.8	12.88	1.9	0.0563	1.7		
817-2.1c	—	0.0564	—	291	0.53	474	8.6	13.11	1.8	0.0564	1.8		
817-3.1c	0.000079	0.057	0.08	163	0.26	466.2	8.8	13.34	1.9	0.0559	2.5		
817-4.1c	0.00032	0.0595	0.44	481	0.51	451.8	8.4	13.79	1.9	0.0549	3.4		
817-5.1c	0.00019	0.0555	—	128	0.63	485.4	9.2	12.85	1.9	0.0528	3		
817-6.1c	0.00011	0.0566	0	134	0.77	475.3	10.1	13.09	2.2	0.055	3		
817-7.1c	0.00018	0.0551	—	164	0.26	442.3	8.3	14.14	1.9	0.0524	4		

(continued)

TABLE 1. SHRIMP U-Th-Pb DATA FOR ZIRCON AND MONAZITE FROM ROCKS OF THE WILMINGTON COMPLEX AND WISSAHICKON FORMATION, AS MAPPED IN DELAWARE (*continued*)

Sample [†] (location)	Measured $\frac{^{204}\text{Pb}}{^{206}\text{Pb}}$	Measured $\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	Common $\frac{^{208}\text{Pb}}{^{206}\text{Pb}}$ (%)	U (ppm)	Th/U	$\frac{^{206}\text{Pb}^{\dagger}}{^{238}\text{U}}$ (Ma)	Error [§] (Ma)	$\frac{^{238}\text{U}^{\dagger}}{^{206}\text{Pb}}$	Error [§] (%)	$\frac{^{207}\text{Pb}^{\dagger}}{^{206}\text{Pb}}$	Error [§] (%)	$\frac{^{207}\text{Pb}^{\dagger, \ddagger}}{^{206}\text{Pb}}$ (Ma)	Error [§] (Ma)
817-8.1c	—	0.0586	0.26	171	0.49	471.4	8.8	13.15	1.9	0.0586	2.2		
817-9.1mix	—	0.0583	0.28	75	0.67	451.1	9.2	13.76	2.1	0.0583	3.4		
817-10.1mix	0.0003	0.0578	0.21	240	0.25	456	8.4	13.69	1.9	0.0535	2.9		
817-11.1c	—	0.0744	—	146	1.43	1096.9	39.2	5.4	3.7	0.0744	1.4	1052	28
817-12.1c	—	0.0545	—	221	0.6	474.6	9.4	13.12	2	0.0545	2		
817-13.1c	—	0.0562	—	174	0.76	472	9	13.17	1.9	0.0562	2.4		
817-14.1c	—	0.0585	0.2	152	0.5	488.3	9.3	12.68	1.9	0.0585	2.7		
817-15.1mix	—	0.0556	—	153	0.5	458.7	8.8	13.57	1.9	0.0556	2.8		
817-16.1c	—	0.0568	0.09	333	0.49	452.9	8.3	13.73	1.9	0.0568	1.7		
817-17.1c	0.00041	0.0573	0.09	77	0.86	474.7	9.5	13.17	2	0.0512	5.2		
G. 43959 (Ardentown Granitic Suite, pyroxene quartz diorite)													
959-1.1c	0.00012	0.0578	0.26	168	0.54	440.1	5.8	14.15	1.4	0.056	3.8		
959-2.1c	0.00022	0.0564	0.11	236	0.48	433	5.4	14.43	1.3	0.0532	2.1		
959-3.1c	0.00068	0.0609	0.69	71	0.44	425.3	11.9	14.74	2.9	0.0509	7.8		
959-4.1c	0.000094	0.0552	—	348	0.82	431.8	5.3	14.46	1.3	0.0538	2.3		
959-5.1c	0.0001	0.0565	0.1	502	0.73	441.4	5.1	14.12	1.2	0.055	1.6		
959-6.1c	—	0.0565	0.12	264	0.63	434.8	5.4	14.28	1.3	0.0588	2.5		
959-7.1c	0.000057	0.0559	—	459	0.9	451.1	5.3	13.81	1.2	0.0551	1.4		
959-8.1c	0.000062	0.0563	0.04	529	0.71	451.9	6.4	13.78	1.4	0.0554	1.5		
959-9.1c	0.00071	0.0564	0.12	503	0.35	429.2	5.9	14.7	1.5	0.0459	8.2		
959-10.1c	0.0012	0.0547	—	352	0.34	431.1	5.2	14.79	1.5	0.0368	19.3		
959-11.1c	0.0001	0.059	0.53	326	1.18	399.8	5	15.58	1.3	0.0575	2.2		
H. 44069 (Wissahickon Formation)													
44069m-1.1c-uz	0.000087	0.0554	—	1426	12.9	436.2	3.0	14.31	0.7	0.0542	1.0		
44069m-2.1c-uz	0.000148	0.0563	0.11	1109	14.6	431.6	3.1	14.46	0.7	0.0542	1.2		
44069m-3.1c-uz	0.000068	0.0549	—	2006	7.4	426.3	3.3	14.65	0.8	0.0539	0.9		
44069m-4.1c-uz	0.000109	0.0562	0.10	1321	10.6	429.3	3.0	14.54	0.7	0.0546	1.0		
44069m-5.1c-uz	0.000108	0.0552	—	1538	10.2	429.8	3.0	14.54	0.7	0.0536	1.0		
44069m-6.1c-sz	0.000047	0.0553	—	1557	9.5	445.2	3.0	14.01	0.7	0.0546	0.6		
44069m-7.1c-uz	0.000091	0.0554	0.02	1429	12.3	423.5	2.9	14.75	0.7	0.0541	1.0		
44069m-8.1c-uz	0.000687	0.0643	1.11	1417	11.0	422.8	3.0	14.77	0.7	0.0542	3.1		
44069m-9.1c-sz	0.000089	0.0556	0.03	1271	13.6	428.5	3.4	14.57	0.8	0.0543	1.1		
44069m-10.1c-uz	0.000075	0.0553	0.01	2545	8.4	422.3	2.8	14.79	0.7	0.0542	0.8		
44069m-11.1c-uz	0.000061	0.0554	0.01	2745	6.7	425.0	2.8	14.69	0.7	0.0545	0.6		
44069m-12.1c-sz	0.000130	0.0565	0.02	2414	10.2	425.3	2.8	14.69	0.7	0.0536	1.0		
44069m-13.1c-uz	0.000060	0.0555	0.02	1650	9.4	425.5	3.6	14.67	0.9	0.0546	0.9		
44069m-14.1i-uz	0.000050	0.0555	0.04	5462	5.1	418.1	2.7	14.93	0.7	0.0547	0.4		
44069m-15.1c-uz	0.000054	0.0557	0.04	2198	6.0	428.0	2.8	14.58	0.7	0.0549	0.6		

[†]Sample names containing the suffix "m" are monazite; all other samples are zircon. Zircon samples 42001 and 43817 analyzed November 1999 on the U.S. Geological Survey–Stanford sensitive high resolution ion microprobe–reverse geometry (SHRIMP-RG[USGS]); samples 43910, 43909-2, 43562, 43753, and 43959 analyzed January 2000 on SHRIMP II at the Research School of Earth Sciences (RSES), Australian National University. Monazite sample 43753 analyzed November 2000 on SHRIMP II (session 1) and April 2003 on SHRIMP-RG (RSES) (session 2); monazite sample 44069 analyzed October 2002 on SHRIMP-RG[USGS]. Abbreviations for zircon: eq—equant, c—core, r—rim, ctz—core, truncated zoning, mix—analysis overlapped two zones. Abbreviations for monazite: c—core, i—intermediate zone, r—rim, oz—oscillatory zoned, sz—spotted zoning, uz—unzoned.

[‡] $^{206}\text{Pb}/^{238}\text{U}$ ages corrected for common Pb using the ^{207}Pb correction method; $^{207}\text{Pb}/^{206}\text{Pb}$ ages corrected for common Pb using the ^{204}Pb correction method. Decay constants from Steiger and Jäger (1977).

[§]1 σ errors.

[#]Radiogenic ratios, corrected for common Pb using the ^{204}Pb correction method, based on the Stacey and Kramers (1975) model.

^{††}Listed only for ages older than 1.0 Ga.

an isobaric interference at mass 204 (Ireland, 1995; Ireland et al., 1999), thereby alleviating the possibility of overcorrection for ^{204}Pb (Rubatto et al., 2001). SHRIMP-RG (RSES and USGS–Stanford) analyses of monazite samples 44069 and 43753 (session 2) did not use energy filtering. For zircon analyses, the primary oxygen ion beam excavated an area of $\sim 25\text{--}35\ \mu\text{m}$ in diameter to a depth of $\sim 0.5\text{--}1\ \mu\text{m}$. For monazite analyses, a smaller beam of $\sim 10\text{--}15\ \mu\text{m}$ was used. Concentrations of U and Th (Tables 1 and 2) are believed to be accurate to about $\pm 20\%$. Zircon standards AS57 (from 1099 Ma gabbroic anorthosite, Duluth Complex; Paces and Miller, 1993) and R33 (419 Ma from monzodiorite, Braintree Complex, Vermont; Black et al., 2004), and monazite standard TM1 (from 1768 Ma pegmatite, Thompson Mine, Manitoba, I.S. Williams, RSES-ANU, 2000, personal commun.) were used to calibrate $^{206}\text{Pb}/^{238}\text{U}$ ages. During the course of this study, we found that isotopic data from monazite sample 44069 were very uniform. To assess the quality of this monazite as a potential SHRIMP standard, this material also was dated by thermal ionization mass spectrometry (TIMS). TIMS Pb/U ages are concordant and are consistent with the SHRIMP results. We then used 44069 monazite as a standard to determine SHRIMP ages for monazite sample 43753 (session 2).

SHRIMP raw data were reduced using the Squid program of Ludwig (2001). Tera-Wasserburg plots (Tera and Wasserburg, 1972) of ^{204}Pb -corrected isotopic data (plotted as 2σ error ellipses) were used only for visual assessment of the data arrays to determine which analyses to include in age calculations. The age of each sample was determined by calculating the weighted average of the $^{206}\text{Pb}/^{238}\text{U}$ ages (Isoplot/Ex program of Ludwig, 2003), using the ^{207}Pb -correction method (Compston et al., 1984). For comparison of geochronologic data acquired in different analytical sessions, the uncertainty in the weighted average age includes propagation of both the 2σ external spot-to-spot error of the standard and the 2σ error of the mean of the standard (as calculated by Squid 1.02; Ludwig, 2001).

For detrital zircon U-Pb data, zircons older than ca. 1.0 Ga were dated using the $^{207}\text{Pb}/^{206}\text{Pb}$ age, whereas younger zircons used the $^{206}\text{Pb}/^{238}\text{U}$ age. Data were screened prior to calculating a relative probability distribution (using Isoplot/Ex; Ludwig, 2003), by plotting on a Tera-Wasserburg concordia plot. Discordant data (i.e., error ellipses that did not intersect the concordia curve) were excluded from the relative probability calculation.

TIMS analyses of monazite (Table 3) were conducted at the Jack Satterly Geochronology Laboratory, Royal Ontario Museum, Toronto.

Monazite was dissolved in 6 N HCl at $100\ ^\circ\text{C}$ for 3 days. Extraction of U and Pb from monazite generally followed the procedure of Krogh (1973), modified by using small anion exchange columns (0.05 mL of resin) that permitted the use of reduced acid reagent volumes. Pb and U were loaded together with silica gel onto outgassed rhenium filaments. The isotopic compositions of Pb and U were measured using a single collector with a Daly pulse counting detector in a solid source VG354 mass spectrometer. Error estimates were calculated by propagating known sources of analytical uncertainty for each analysis based on long-term replicate measurements of the standards SRM982 and CBNM72–6, and uncertainties in the isotopic composition and amount of laboratory blank and initial Pb. Decay constants used were those of Jaffey et al. (1971).

All age errors quoted in the text and in Table 4 (summary) are given at the 95% confidence interval. Error ellipses and error bars in the concordia diagrams and insets are plotted at 2σ .

Sample Descriptions and Results

Because several of the rocks analyzed in this study have been metamorphosed to upper amphibolite or granulite facies, observation and interpretation of zircon morphology is critical for understanding the isotopic data. For example, euhedral, concentric oscillatory zoning (in CL) and euhedral, prismatic external morphology are generally regarded as evidence for crystallization of zircon from a magma. In contrast, zircon that forms in a high-grade metamorphic environment tends to have patchwork zoning and multifaceted, equant, tabular external morphology (van Breemen et al., 1986; Vavra et al., 1999; Kröner et al., 2000; Pidgeon et al., 2000). At very high metamorphic grade, these distinctions of origin become less obvious, particularly if metamorphic temperatures become high enough to cause partial melting. Under anatexis conditions, it becomes a semantic issue as to whether a zircon is considered to be igneous or metamorphic in origin. Zircons that have equant morphology occur in both igneous and sedimentary rocks and can superficially resemble equant zircons of metamorphic origin. However, equant igneous zircons can be distinguished from metamorphic zircons by simpler crystallography (lower-order crystal faces) and concentric oscillatory zoning in CL. Equant zircons of detrital origin in sedimentary rocks tend to be spherical (not discoidal), usually have pitted or frosted, rounded surfaces (not flat faces), and truncated zoning in CL. As will be shown below, several rock samples from the Wilmington Complex contain equant zircon of igneous, metamorphic

and (or) sedimentary origin. Careful inspection of zircon morphology, in conjunction with the geochronologic and zoning data, is critical for determining the history of the rock.

Brandywine Blue Gneiss

Two populations of zircon (equant and elongate) are found in the Brandywine Blue Gneiss (sample 43910). Elongate grains have length-to-width ratios (l/w) of $\sim 3\text{--}5$, whereas equant grains have l/w of $1\text{--}2$ (Figs. 3A–C). In CL, all of the zircons, regardless of morphology, contain oscillatory zoning; some grains also have overgrowths of varying thickness, which contain patchwork zoning or are unzoned, and usually are darker in CL than the adjacent core. SHRIMP data fall into two age groups, 476 ± 6 and 428 ± 4 Ma (Fig. 4A), each of which includes both equant and elongate zircons. A plot of Th/U versus age for zircons from the Brandywine Blue Gneiss clearly distinguishes the two age groups (Fig. 5). Older grains have relatively low Th/U of $\sim 0.15\text{--}0.35$, whereas younger grains have relatively high Th/U of $\sim 0.35\text{--}0.75$; all grains have Th/U typical of igneous zircons. We conclude that the Brandywine Blue Gneiss protolith was emplaced at ca. 476 ± 6 Ma and was metamorphosed to granulite facies ~ 50 m.y. later, causing the growth of new zircon at 428 ± 4 Ma. A few analyses plot slightly outside of these two age groupings (Fig. 4A). Four grains with intermediate ages between ca. 476 and 428 Ma (not included in either weighted average calculation) have Th/U identical to analyses of the older age group; the two youngest grains, also excluded from the weighted average calculation, have identical Th/U to those of the younger age group. Possible explanations for the slightly younger ages are Pb loss, mixing of two (or more) age components, or growth of new zircon at an intermediate age of ca. 450 Ma. The Th/U data suggest that minor Pb loss is the most likely cause for the intermediate and young ages.

Rockford Park Gneiss

Two populations of zircon (equant and elongate) occur in the sample of felsic gneiss from the Rockford Park Gneiss (sample 43909-2). Elongate zircons have l/w of $3\text{--}5$ and contain concentric oscillatory zoning in CL. Nearly all of the elongate zircons contain overgrowths that are dark in CL (i.e., relatively high U; Table 1), generally occurring as thin rinds around the grains, although a few overgrowths are quite broad and nearly as extensive as the cores (Fig. 3D). The weighted average age for 12 oscillatory zoned cores, plus one rim, all from elongate grains, is 476 ± 4 Ma (Fig. 4B). This age is interpreted as the time of crystallization

TABLE 2. SHRIMP U-Th-Pb DATA FOR DETRITAL ZIRCON FROM THE WINDY HILLS GNEISS AND WISSAHICKON FORMATION, AS MAPPED IN DELAWARE

Sample [†]	Measured $\frac{^{204}\text{Pb}}{^{206}\text{Pb}}$	Measured $\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	Common $\frac{^{206}\text{Pb}}{^{206}\text{Pb}}$ (%)	U (ppm)	Th/U	$\frac{^{206}\text{Pb}^{\dagger}}{^{238}\text{U}}$ (Ma)	Error [§] (Ma)	$\frac{^{207}\text{Pb}^{\dagger}}{^{206}\text{Pb}}$ (Ma)	Error [§] (Ma)	$\frac{^{238}\text{U}^{\dagger}}{^{206}\text{Pb}}$ (%)	Error [§] (%)	$\frac{^{207}\text{Pb}^{\dagger}}{^{206}\text{Pb}}$ (%)	Error [§] (%)
C. 43592 (Windy Hills Gneiss detrital zircons)													
562-1.1	0.00039	0.0673	1.15	246	0.89	526.5	8.5	660	78	11.7	1.7	0.0616	3.6
562-2.1	—	0.0844	—	319	0.4	1331.3	17.1	1301	19	4.37	1.3	0.0844	1
562-3.1	0.0006	0.0635	0.86	103	0.58	473.9	8.4	398	105	13.14	1.8	0.0546	4.7
562-4.1	—	0.0739	—	250	0.48	1050.4	14	1040	28	5.65	1.4	0.0739	1.4
562-5.1	0.00012	0.0965	0.09	204	0.29	1542.9	20.2	1524	26	3.7	1.3	0.0948	1.4
562-6.1	0.00014	0.0746	0.05	372	0.05	1045.4	12.9	1002	27	5.69	1.3	0.0726	1.3
562-7.1	0.00028	0.0818	1.45	124	0.58	919.7	13.2	1143	57	6.46	1.5	0.0778	2.9
562-8.1	0.000078	0.0725	—	159	1.62	1069.5	14.8	967	40	5.57	1.4	0.0713	2
562-9.1	0.00007	0.0729	0.24	261	0.11	954.7	12.6	985	33	6.26	1.4	0.072	1.6
562-10.1	0.0004	0.0819	0.42	173	0.52	1156.6	15.6	1101	59	5.1	1.4	0.0762	2.9
562-11.1	0.00012	0.0777	0.41	67	1.18	1048.7	16.4	1095	56	5.65	1.6	0.076	2.8
562-12.1	0.00016	0.0796	0.03	249	0.28	1180.6	14.9	1131	33	4.99	1.3	0.0774	1.7
562-13.1	0.000017	0.1084	0.61	411	0.16	1680.9	21	1768	19	3.34	1.3	0.1081	1
562-14.1	0.0001	0.0839	0.41	73	0.9	1207.7	18.2	1256	42	4.84	1.6	0.0824	2.1
562-15.1	0.000053	0.1111	1.49	274	0.71	1591.7	20.1	1806	15	3.52	1.3	0.1104	0.8
562-16.1	0.00017	0.0792	0.01	268	0.63	1176.4	15.1	1115	38	5.01	1.3	0.0768	1.9
562-17.1	0.00015	0.0756	0.18	288	0.37	1043.2	13.2	1027	38	5.7	1.3	0.0735	1.9
562-18.1	0.000041	0.0811	—	487	0.3	1240.4	15	1211	17	4.72	1.3	0.0806	0.8
562-19.1	0.00039	0.061	0.55	415	0.08	474.7	6.5	424	82	13.11	1.4	0.0553	3.7
562-20.1	0.000089	0.0799	0.28	281	0.62	1135.9	14.2	1163	24	5.18	1.3	0.0786	1.2
562-21.1	0.00027	0.0816	1.24	37	0.94	966.4	18.1	1141	73	6.14	1.9	0.0778	3.7
562-22.1	0.000031	0.0782	—	779	0.34	1169.2	14.3	1141	14	5.03	1.3	0.0778	0.7
562-23.1	0.000044	0.0907	0.39	197	0.28	1367.8	18.5	1427	24	4.22	1.4	0.0901	1.3
562-24.1	—	0.1063	0.35	153	0.85	1683.7	24.1	1738	26	3.34	1.5	0.1063	1.4
562-25.1	0.000078	0.0735	—	402	0.44	1044.9	12.8	998	26	5.69	1.3	0.0724	1.3
562-26.1	0.00076	0.0635	0.8	170	0.42	491.6	7	299	199	12.69	1.5	0.0523	8.7
562-27.1	0.000076	0.1195	0.31	172	0.39	1906.5	29	1933	18	2.9	1.5	0.1185	1
562-28.1	0.00017	0.0831	0.35	141	0.66	1201.5	16.9	1216	31	4.88	1.4	0.0808	1.6
562-29.1	0.000068	0.0619	0.14	835	0.93	629.9	7.6	637	36	9.74	1.2	0.0609	1.7
562-30.1	0.00015	0.0574	0.13	603	0.19	466.9	5.8	419	51	13.33	1.3	0.0552	2.3
562-31.1	0.00022	0.0738	0.84	118	1.02	833.6	11.8	946	89	7.21	1.5	0.0706	4.4
562-32.1	0.00012	0.1024	1	362	0.79	1504.7	18.5	1639	17	3.77	1.3	0.1008	0.9
562-33.1	0.0014	0.0648	1.04	110	0.79	465.9	7.3	-101	422	13.54	1.8	0.0442	17.2
562-34.1	—	0.0784	—	283	0.65	1163.5	15.6	1158	21	5.06	1.4	0.0784	1.1
562-35.1	0.000085	0.0702	0.58	130	0.43	785	11.7	897	49	7.69	1.5	0.0689	2.4
562-36.1	0.00011	0.0842	—	156	0.46	1313.4	18.1	1260	29	4.44	1.4	0.0826	1.5
562-37.1	0.000039	0.0797	0.47	324	0.21	1088.4	13.8	1176	24	5.42	1.3	0.0792	1.2
562-38.1	0.000031	0.0868	0.79	341	0.27	1201.1	14.8	1347	19	4.85	1.3	0.0864	1
562-39.1	0.000073	0.0875	0.57	211	0.45	1262.9	16.2	1350	28	4.6	1.3	0.0865	1.4
562-40.1	0.00032	0.0699	0.93	240	0.29	685.6	8.9	786	67	8.88	1.3	0.0654	3.2
H. 44069 (Wissahickon Formation psammitic gneiss)													
069-1.1E-core	0.00012	.0845	0.53	252	0.23	1197.5	15.1	1265	25	4.88	1.3	.0828	1.3
069-2.1E-core	0.00018	.0759	0.71	188	0.45	927.7	13.5	1020	61	6.44	1.5	.0732	3.0
069-3.1E-rim	0.00000	.0714	0.60	348	0.08	820.8	10.2	968	24	7.32	1.3	.0714	1.2
069-4.1E-core	0.00034	.0788	0.86	28	0.84	976.3	20.5	1041	100	6.10	2.2	.0740	5.0
069-5.1E-rim	0.00017	.0732	0.33	253	0.13	942.1	13.0	950	44	6.35	1.4	.0707	2.1
069-6.1E-rim	0.00003	.0732	0.27	472	0.09	954.8	11.7	1005	18	6.25	1.3	.0727	0.9
069-7.1E-rim	0.00023	.0699	—	340	0.29	1168.0	20.0	825	37	5.11	1.8	.0666	1.8
069-1.1	0.00013	.0746	0.16	236	0.06	1020.1	13.8	1006	32	5.84	1.4	.0727	1.6
069-2.1	0.00050	.0738	0.45	68	0.45	929.5	16.5	828	143	6.47	1.9	.0667	6.9

(continued)

TABLE 2. SHRIMP U-Th-Pb DATA FOR DETRITAL ZIRCON FROM THE WINDY HILLS GNEISS AND WISSAHICKON FORMATION, AS MAPPED IN DELAWARE (*continued*)

Sample [†]	Measured $\frac{^{204}\text{Pb}}{^{206}\text{Pb}}$	Measured $\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	Common $\frac{^{206}\text{Pb}}{^{206}\text{Pb}}$ (%)	U (ppm)	Th/U	$\frac{^{206}\text{Pb}^{\ddagger}}{^{238}\text{U}}$ (Ma)	Error [§] (Ma)	$\frac{^{207}\text{Pb}^{\ddagger}}{^{206}\text{Pb}}$ (Ma)	Error [§] (Ma)	$\frac{^{238}\text{U}^{\ddagger}}{^{206}\text{Pb}}$ (Ma)	Error [§] (%)	$\frac{^{207}\text{Pb}^{\ddagger}}{^{206}\text{Pb}}$ (%)	Error [§] (%)
069-3.1	0.00003	.0746	0.31	374	0.47	988.3	12.6	1046	24	6.02	1.3	.0742	1.2
069-4.1	0.00007	.0803	0.44	385	0.29	1110.6	14.1	1178	26	5.30	1.3	.0792	1.3
069-5.1	0.00059	.0835	1.08	87	0.34	1053.6	17.0	1074	106	5.63	1.7	.0752	5.3
069-6.1	0.00062	.0755	0.60	79	0.38	942.4	15.2	824	146	6.38	1.7	.0666	7.0
069-7.1	0.00005	.0743	0.07	363	0.06	1035.5	13.3	1032	24	5.74	1.3	.0737	1.2
069-8.1	0.00088	.0821	1.07	64	0.23	1017.8	17.1	912	177	5.87	1.8	.0694	8.6
069-9.1	0.00039	.1062	0.75	61	0.43	1619.0	26.2	1639	45	3.50	1.7	.1008	2.4
069-10.1	0.00000	.0773	0.29	182	0.48	1065.1	14.2	1128	34	5.55	1.4	.0773	1.7
069-11.1	0.00018	.0903	0.55	124	0.55	1328.6	18.5	1379	35	4.36	1.4	.0878	1.8
069-12.1	0.00008	.1075	0.50	476	0.18	1682.2	20.9	1740	12	3.34	1.3	.1065	0.7
069-13.1	0.00024	.0800	0.78	60	0.36	1029.0	18.5	1112	75	5.76	1.9	.0766	3.8
069-14.1	0.00009	.0734	—	239	0.30	1061.3	13.7	988	33	5.61	1.3	.0721	1.6
069-15.1	0.00042	.0708	—	141	0.44	1000.8	14.7	766	87	6.01	1.5	.0648	4.1
069-16.1	0.00019	.0842	1.37	132	0.23	1008.8	14.1	1234	42	5.84	1.4	.0815	2.2
069-17.1	0.00064	.0737	0.18	57	0.72	992.7	18.1	758	183	6.06	2.0	.0645	8.7
069-18.1	0.00000	.0833	0.25	111	0.60	1225.6	17.4	1277	34	4.76	1.5	.0833	1.8
069-19.1	0.00002	.0753	0.44	789	0.34	975.4	11.8	1068	21	6.10	1.2	.0750	1.1
069-20.1	0.00012	.0765	0.15	284	0.21	1074.3	14.1	1063	34	5.52	1.4	.0748	1.7
069-21.1	0.00012	.0764	0.26	193	0.21	1047.6	13.8	1061	39	5.66	1.4	.0747	1.9
069-22.1	0.00009	.0782	—	181	0.52	1238.5	17.1	1119	39	4.75	1.4	.0769	2.0
069-23.1	0.00000	.0807	0.27	189	0.34	1157.5	15.3	1214	28	5.07	1.4	.0807	1.4
069-24.1	0.00027	.0707	—	129	0.30	959.2	13.5	834	64	6.26	1.5	.0669	3.0
069-25.1	0.00005	.0767	0.09	682	0.26	1093.2	13.3	1097	18	5.41	1.3	.0761	0.9
069-26.1	0.00015	.0744	1.15	100	0.31	772.3	11.6	993	64	7.79	1.5	.0723	3.2
069-27.1	0.00005	.0788	0.12	480	0.21	1142.9	14.7	1151	20	5.15	1.3	.0782	1.0
069-28.1	0.00013	.0681	0.52	518	0.15	735.3	9.2	814	27	8.25	1.3	.0662	1.3
069-29.1	0.00017	.0761	0.24	150	0.21	1044.2	14.2	1033	53	5.69	1.4	.0737	2.6
069-30.1	0.00010	.0794	0.58	80	1.15	1054.3	16.0	1145	50	5.60	1.6	.0779	2.5
069-31.1	0.00007	.0734	—	234	0.40	1043.9	20.4	996	28	5.70	2.0	.0723	1.4
069-32.1	0.00032	.0849	—	166	0.48	1323.5	17.5	1206	53	4.41	1.4	.0804	2.7
069-33.1	0.00014	.0765	0.21	292	0.45	1061.9	13.4	1056	35	5.59	1.3	.0745	1.7
069-34.1	0.00003	.0724	—	177	0.40	1026.4	13.7	986	47	5.80	1.4	.0720	2.3
069-35.1	0.00081	.0786	0.82	66	0.08	979.4	15.5	839	159	6.13	1.7	.0670	7.6
069-36.1	0.00045	.0732	0.28	121	0.38	954.3	13.4	828	89	6.30	1.5	.0667	4.3
069-37.1	0.00006	.1067	0.41	308	0.70	1680.0	22.4	1729	14	3.35	1.4	.1059	0.8
069-38.1	0.00023	.0771	0.27	161	0.50	1065.0	14.3	1036	60	5.57	1.4	.0738	3.0
069-39.1	0.00028	.0727	0.33	239	0.38	926.3	12.3	889	35	6.48	1.4	.0687	1.7
069-40.1	0.00024	.0747	0.46	158	0.68	954.2	12.9	965	61	6.27	1.4	.0713	3.0
069-41.1	0.00014	.0772	0.34	166	0.21	1049.7	14.0	1074	55	5.65	1.4	.0752	2.7
069-42.1	0.00004	.0755	0.13	456	0.16	1052.8	12.8	1067	20	5.63	1.3	.0750	1.0
069-43.1	0.00018	.0707	—	90	0.51	976.5	14.4	871	54	6.14	1.5	.0681	2.6
069-44.1	0.00016	.0881	0.87	191	0.47	1215.7	16.0	1335	47	4.79	1.3	.0859	2.4
069-45.1	0.00000	.0740	0.42	130	0.52	943.9	13.1	1043	37	6.31	1.4	.0740	1.9
069-46.1	0.00029	.0728	—	72	0.21	1032.3	15.8	888	84	5.79	1.6	.0687	4.0
069-47.1	0.00008	.0731	—	280	0.22	1028.6	13.1	984	43	5.79	1.3	.0719	2.1
069-48.1	0.00012	.0800	0.04	181	0.48	1189.1	15.7	1155	32	4.94	1.4	.0783	1.6
069-49.1	0.00015	.0722	0.02	272	0.53	985.0	12.6	930	40	6.07	1.3	.0701	1.9
069-50.1	0.00071	.0866	2.08	81	0.44	907.8	14.5	1108	155	6.56	1.7	.0765	7.7

[†]All samples analyzed on the U.S. Geological Survey–Stanford sensitive high resolution ion microprobe (SHRIMP-RG) in June 2002. Wissahickon sample names containing "E-" are elongate grains.

[‡] $^{206}\text{Pb}/^{238}\text{U}$ ages corrected for common Pb using the ^{207}Pb correction method; $^{207}\text{Pb}/^{206}\text{Pb}$ ages, $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{238}\text{U}/^{206}\text{Pb}$ corrected for common Pb using the ^{204}Pb correction method.

[§]1 σ error.

TABLE 3. U-Pb THERMAL IONIZATION MASS SPECTROMETRY (TIMS) DATA FOR MONAZITE SAMPLE 44069 (WISSAHICKON FORMATION)

Sample [†]	Weight (mg)	U (ppm)	Th/U	Common Pb (pg)	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$ [‡] (Ma)	Error [§] (Ma)	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$ [‡] (Ma)	Error [§] (Ma)	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$ [‡] (Ma)	Error [§] (Ma)	rho
m1-w, eq	0.006	2901	9.6	6.8	11212	0.06815	0.5200	425.0	1.0	425.1	1.0	426.0	3.0	0.886
m2-fr	0.004	2036	13.2	1.9	18390	0.06806	0.5193	424.5	1.0	424.7	0.9	425.8	2.6	0.907
m3-h	0.009	2035	12.2	11.4	7025	0.06821	0.5201	425.3	1.3	425.2	1.3	424.5	4.3	0.855
m4-w, eq	0.007	2924	9.0	4.3	20965	0.06813	0.5199	424.9	0.9	425.1	0.9	426.0	2.6	0.898
m5-fr	0.004	2799	11.1	7.8	6283	0.06821	0.5199	425.4	1.1	425.1	1.2	423.7	5.0	0.747
m6-fr	0.007	1684	13.4	4.6	11230	0.06822	0.5205	425.4	1.0	425.5	1.0	426.1	3.4	0.841
m7-h	0.006	2134	11.5	6.2	9094	0.06813	0.5198	424.9	1.1	425.0	1.1	425.5	3.5	0.872
m8-w, eq	0.003	2167	11.6	2.3	12628	0.06790	0.5177	423.5	1.4	423.6	1.2	424.0	4.2	0.856
m9-fr, el	0.003	4099	10.4	2.2	24277	0.06795	0.5181	423.8	1.8	423.9	1.5	424.3	2.7	0.963
Weighted average [#]								424.86 ± 0.36		424.89 ± 0.35		425.3 ± 1.1		
MSWD ^{††}								1.06		1.10		0.30		
Concordia age ^{‡‡}								424.86 ± 0.50						

[†]All analyses made on single clear, yellow grains. Abbreviations: w—whole grain, eq—equant, fr—fragment, h—half grain, el—elongate.

[‡]Blank and fractionation corrected. Total common Pb of less than or equal to 10 pg corrected using the blank Pb isotopic composition. Common Pb >10 pg is assumed to be initial Pb and was corrected using Stacey and Kramers (1975) Pb evolution model.

[§]2σ errors.

[#]Does not include uncertainty in decay constants.

^{††}Mean square of the weighted deviates (MSWD).

^{‡‡}Calculated by Isoplot/Ex (Ludwig, 2001) following algorithm of Ludwig (1998); includes decay constant errors.

TABLE 4. SUMMARY OF U-Pb GEOCHRONOLOGY RESULTS

Sample number	Lithologic unit	Zircon morphology [†]	Age [‡] (Ma)	Origin [§]
43910	Brandywine	Elongate (co) + equant (co)	476 ± 6	ign
	Blue Gneiss	Elongate (co) + equant (co)	428 ± 4	meta
43909-2	Rockford	Elongate (co)	506, 507	inh
	Park Gneiss	Elongate (co)	476 ± 4	ign
		Equant (pw)	432 ± 6	meta
43562	Windy Hills	Elongate (tco) + equant (tco)	527–1933	det
	Gneiss	Elongate (co)	482 ± 4	ign
43753	Faulkland	Elongate (co)	772, 1106	det
	Gneiss	Elongate (co) + equant (co)	481 ± 4	ign
		<i>monazite (uz, pw, co)</i>	447 ± 3	meta
			429 ± 2	meta
			411 ± 3	meta
			398 ± 3	meta
42001	Christianstead	Elongate (co)	511	inh
	Gneiss	Elongate (co)	483 ± 7	ign
43817	Barley Mill	Elongate (co)	476 ± 8	ign
	Gneiss			
43959	Ardentown	Elongate (co)	434 ± 5	ign
	Granitic Suite			
44069	Wissahickon Fm. metapsammite	Elongate (tco) + equant (tco)	735–1740	det
		<i>monazite (uz)</i>	426 ± 3	meta
		Monazite (uz) [TIMS data]	424.9 ± 0.4	meta

[†]Letters in parentheses describe zoning patterns as shown by cathodoluminescence (for zircon) or backscattered electrons (for monazite). Elongate = length-to-width ratio >2, equant = length-to-width ratio <2. Abbreviations: co—concentric oscillatory zoning, pw—patchwork zoning, tco—truncated concentric oscillatory zoning, uz—unzoned.

[‡]SHRIMP ages with 2σ uncertainties are weighted average of individual $^{206}\text{Pb}/^{238}\text{U}$ ages. Ages without uncertainties are from individual grains; $^{207}\text{Pb}/^{206}\text{Pb}$ used for ages older than 1.0 Ga, $^{206}\text{Pb}/^{238}\text{U}$ used for ages younger than 1.0 Ga. Sample 44069 monazite ages determined by SHRIMP and thermal ionization mass spectrometry (TIMS).

[§]Abbreviations: ign—igneous, meta—metamorphic, det—detrital, inh—inherited core.

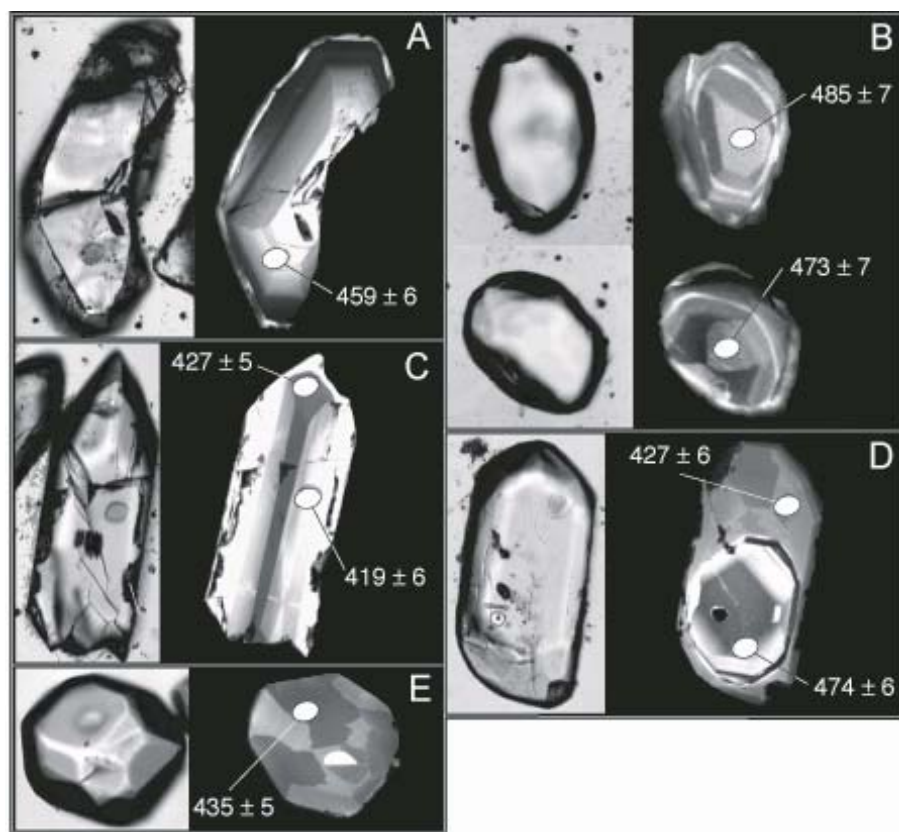


Figure 3. Transmitted-light (left side) and cathodoluminescence (right side) images of representative zircons from granulite-grade rocks of the Wilmington Complex. In this, and all subsequent images of zircons, long axis of SHRIMP analytical spot is $\sim 25 \mu\text{m}$. Errors on $^{206}\text{Pb}/^{238}\text{U}$ ages, shown in this and subsequent figures, are $\pm 1\sigma$. (A) Elongate igneous zircon, Brandywine Blue Gneiss (sample 43910). (B) Equant igneous zircons, Brandywine Blue Gneiss (sample 43910). (C) Elongate metamorphic zircon, Brandywine Blue Gneiss (sample 43910). (D) Elongate zircon with igneous core and broad metamorphic overgrowth, Rockford Park Gneiss (sample 43909-2). (E) Equant metamorphic zircon, Rockford Park Gneiss (sample 43909-2).

of the protolith of the felsic gneiss, probably a volcanic rock, based on field appearance and occurrence. Four analyses of oscillatory zoned cores were excluded from the weighted average calculation: analyses 909-9.1 and 909-17.1 have older ages of 506 and 507 Ma, respectively, and analyses 909-3.1 and 909-14.4 have slightly younger ages of 457 and 451 Ma, respectively (Table 1). Equant zircons have l/w of 1–2 and mostly contain patchwork zoning in CL (Fig. 3E); a few equant grains with oscillatory zoning also were found. The age derived from analysis of 6 equant grains with patchwork zoning, 2 elongate grains with irregular internal zoning in CL (analyses 909-1.1 and 909-5.1), and one dark rim on an elongate grain (analysis 909-4.2, Table 1) is 432 ± 6 Ma (Fig. 4B). We interpret this age to be the time of granulite facies metamorphism.

Windy Hills Gneiss

On the basis of interlayered felsic and mafic lithologies, the Windy Hills Gneiss is interpreted as an altered metavolcanic rock. Two populations of zircon (equant and elongate) occur in sample 43562 (now cordierite-gedrite gneiss). Elongate zircons have l/w of 3–5 and generally have rounded apices and edges (Fig. 6A). Most are broken and have only one pyramidal termination, so that oscillatory zoning in CL is not fully concentric. Only a few grains have very thin, dark (in CL) overgrowths. Twelve cores of elongate grains, plus one dark overgrowth, yield a weighted average age of 482 ± 4 Ma (Fig. 7A), interpreted as the time of extrusion of the protolith of the gneiss.

Equant zircons (l/w of 1–2) also were selected for analysis because we assumed, on the basis of external morphology as viewed by binocular

microscope, that these grains were metamorphic in origin (similar to equant zircons from the Rockford Park Gneiss, described above). When viewed under higher magnification in transmitted light, the equant grains displayed irregular, abraded surfaces; CL imaging showed a wide variety of zoning patterns, including oscillatory and patchwork, both of which typically were truncated (Fig. 6B). The external morphology and zoning patterns suggest that these grains are detrital, rather than metamorphic, in origin. The occurrence of detrital zircons within the Windy Hills Gneiss supports our interpretation that the protolith was an extrusive rock. Fourteen analyses yield ages ranging from ca. 545 to 1851 Ma (Table 1), older than the interpreted extrusion age of the Windy Hills Gneiss protolith.

To better understand the age distribution of detrital zircons in the Windy Hills Gneiss, we analyzed 40 additional grains that appeared to be detrital in origin (Fig. 8A). Five relatively young zircons yield a weighted average of 474 ± 13 Ma, within uncertainty of the 482 ± 4 Ma igneous age of the gneiss protolith. The other grains range in age from ca. 527 to 1933 Ma, with major peaks at ca. 1.05, 1.15, and 1.22 Ga. Detrital zircons older than ca. 900 Ma have an age distribution similar to detrital zircons from the Wissahickon Formation, indicating similar provenance for both lithologies.

Because most of the zircons in the Windy Hills Gneiss clearly are detrital in origin (on the basis of age and morphology), it is possible that all zircons in this sample are detrital. In this case, the age of the elongate population (482 ± 4 Ma) would be the maximum age for the sample.

Faulkland Gneiss

Both equant and elongate zircons occur in the sample of felsic gneiss from the Faulkland Gneiss (sample 43753). Elongate zircons have l/w of 2–5 and show fine concentric, euhedral, oscillatory zoning in CL (Fig. 6C). Many of the grains contain numerous inclusions and are cracked, making it difficult to find pristine areas for isotopic analysis. A small percentage of the grains have thin, dark (in CL) overgrowths. Equant zircons have l/w of 1–2 and concentric oscillatory zoning in CL, indicative of igneous origin (Fig. 6D). The weighted average age for 11 of 16 elongate grains, plus 3 equant grains, is 481 ± 4 Ma (Fig. 7B), interpreted as the time of extrusion of the protolith of this metavolcanic rock. Data from two elongate grains (analyses 753-9.1 and 753-15.1, Table 1) have older ages of 1169 and 773 Ma, respectively. These zircons are similar in morphology and zoning to other grains in the elongate population, but, on the basis of age, are interpreted to be detrital or inherited in origin. Analyses 753-2.1, 753-8.1,

and 753-10.1 (Table 1), in fine oscillatory zoned zircon, yield somewhat younger ages, perhaps due to minor Pb loss through undetected cracks or imperfections.

Monazite from the Faulkland Gneiss (sample 43753) was analyzed to determine the time(s) of metamorphism of the Faulkland Gneiss, and by extension, the entire Wilmington Complex. The monazite grains are anhedral, pale to bright orangish-yellow, and contain a wide variety of zoning patterns (shown in BSE; Fig. 9A–C). Many grains have oscillatory zoned cores overgrown by one or more unzoned mantles or rims. Other common zoning patterns are patchwork or spotted appearance. Some cores are completely unzoned, overgrown by lighter or darker, unzoned mantles or rims. Although the zoning patterns in BSE are due to compositional variations, this zoning shows crosscutting relationships, which suggest that the monazite grew episodically, as has been documented elsewhere (cf. Fanning and Aleinikoff, 1998; Hawkins and Bowering, 1999; Pyle and Spear, 2003). $^{206}\text{Pb}/^{238}\text{U}$ ages range from ca. 392 to 450 Ma. Results from session 2 yield slightly younger $^{207}\text{Pb}/^{206}\text{Pb}$ ages than data obtained during session 1, perhaps due to overcorrection for common Pb at mass 204, because energy filtering was not used. This effect does not affect the age calculation because $^{206}\text{Pb}/^{238}\text{U}$ ages utilize the ^{207}Pb -correction method for determining the common Pb component of each analysis.

Interpreting the wide range of monazite ages is difficult. Only grain 12 has evidence for a clearly older core (analysis 12.1, session 1; 441.7 ± 6.9 Ma) and younger rim (analysis 12.2; 414.8 ± 6.7 Ma) (Table 1, Fig. 9). The two oldest ages, 448 and 449 Ma (analyses 753m-3.2 and 753m-11.1, session 2, respectively) are from relatively dark interior zones surrounded by lighter-colored mantles. Also, the youngest age found (388 Ma; analysis 753m-8.2, session 2) occurs at the edge of a grain in a light-colored overgrowth. However, most zoning distinctions within grains are subtle and irregular and offer no basis for subdividing the ages into groups. Thus, we used the unmix routine of Isoplot/Ex (Ludwig, 2003) to determine the statistical likelihood of multiple ages. $^{206}\text{Pb}/^{238}\text{U}$ ages for each analytical session were unmixed separately (Fig. 10). Isoplot/Ex determined that the most likely deconvolution results in the same four age groups for both data sets. These ages and uncertainties were combined to yield composite ages of 398 ± 3 , 411 ± 3 , 429 ± 2 , and 447 ± 4 Ma (Fig. 10).

Christianstead Gneiss

Zircons from the Christianstead Gneiss (sample 42001) are well preserved. They are euhedral

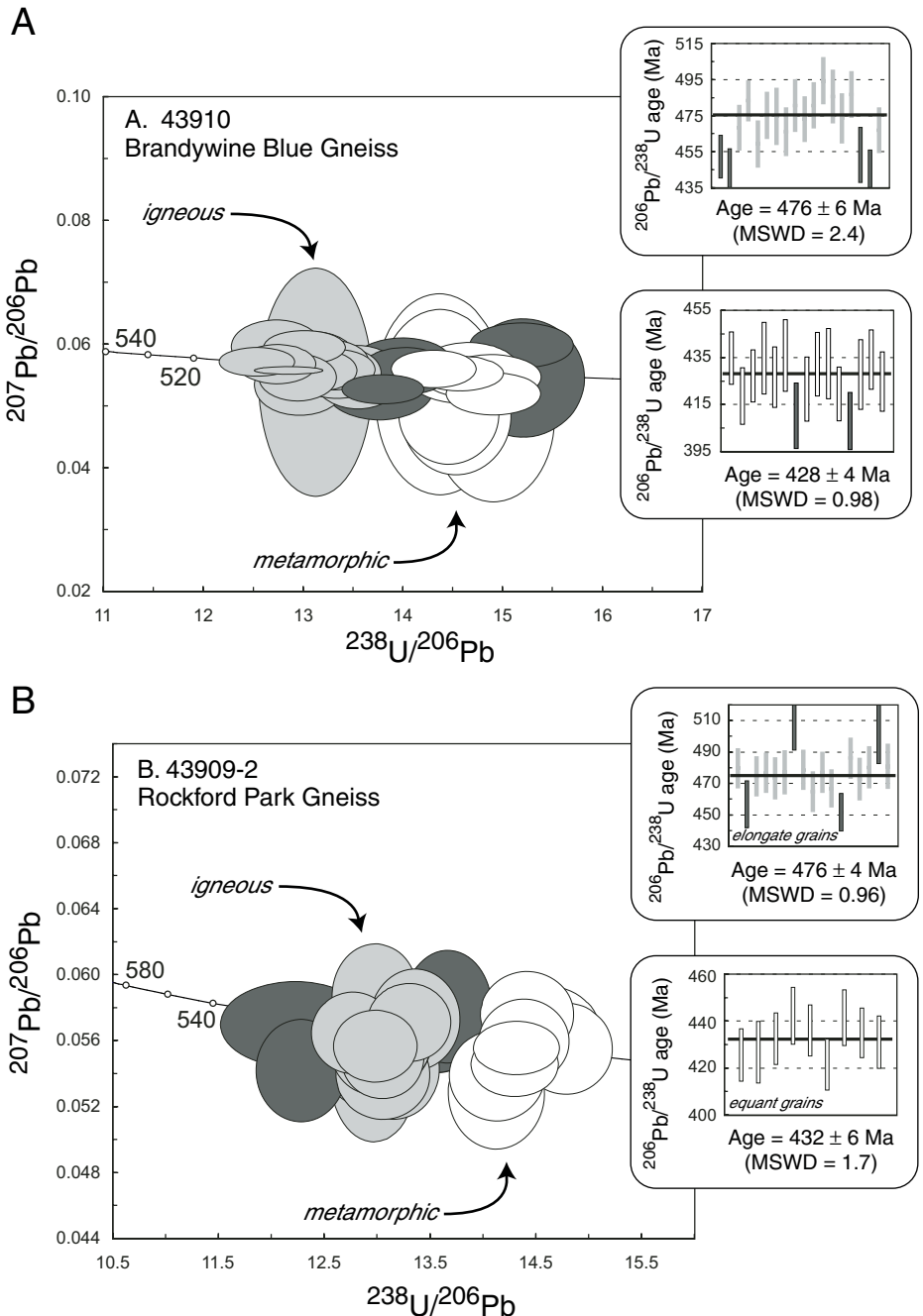


Figure 4. Tera-Wasserburg concordia diagrams and weighted average plots of zircon SHRIMP data. The $^{206}\text{Pb}/^{238}\text{U}$ ages from analyses shown as light-gray error ellipses were used to calculate igneous ages. Metamorphic ages were calculated using $^{206}\text{Pb}/^{238}\text{U}$ ages from analyses shown as white error ellipses. Data from analyses shown as dark-gray error ellipses have been excluded from age calculations (see text for further explanation). (A) Brandywine Blue Gneiss. (B) Rockford Park Gneiss. Here, and in subsequent figures, MSWD—mean square of weighted deviates.

with l/w of 3–6 and contain fine concentric oscillatory zoning (Fig. 11A). No overgrowths were observed. Many of the grains contain fluid and mineral inclusions. Seventeen of 18 analyses yield a weighted average age of 483 ± 7 Ma (Fig. 12A), interpreted as the time of emplace-

ment of the granodiorite gneiss protolith. Analysis 001-3.1 has a slightly older age of 510 Ma.

Barley Mill Gneiss

Zircons from the Barley Mill Gneiss (sample 43817) are elongate and tabular. They have l/w

of 4–6, and all grains contain numerous acicular or round mineral inclusions (Fig. 11B). Many grains are cracked and have thin to thick, dark (in CL) overgrowths. One prismatic grain (analysis 817-11.1) has an age of 1097 Ma. Ten of 16 tabular grains yield a weighted average age of 476 ± 8 Ma (Fig. 12B). Analyses excluded from the calculation have significantly younger ages (Table 1) due to either Pb loss through imperfections, such as cracks, or because the analysis inadvertently overlapped two age components (as revealed by CL images). No rims were analyzed.

Ardentown Granitic Suite Pyroxene Quartz Diorite

Zircons from pyroxene quartz diorite of the Ardentown Granitic Suite of the Arden Plutonic Supersuite (sample 43959) are poorly preserved. Most grains are cracked and many contain fluid inclusion tubes oriented parallel to the *c* axis (Fig. 11C). CL images show dark oscillatory zoning patterns indicative of relatively high U content (Table 1). Eight of 11 analyses yield a weighted average age of 434 ± 5 Ma (Fig. 13), interpreted as the time of emplacement of the quartz diorite (and probably the rest of the Arden Plutonic Supersuite). Three analyses were excluded from the age calculation. Analyses 959-7.1 and 959-8.1 are slightly older at 451 and 452 Ma, respectively; analysis 959-11.1 has a younger age of 400 Ma (Table 1).

Wissahickon Formation

Zircon and monazite were extracted from a psammitic gneiss layer (sample 44069) that occurs within a metasedimentary sequence composed primarily of pelitic paragneiss and amphibolite. This sample was collected near Yorklyn, Delaware, ~8 km northwest across strike from rocks of the Wilmington Complex (Fig. 2). The metasedimentary rocks at this locality are interpreted (by W. Schenck, 2002, personal commun.) as occurring at the stratigraphic bottom of the broad belt of Wissahickon Formation that trends northeast across northern Delaware and southeastern Pennsylvania. There is considerable controversy concerning the correlation of rocks called Wissahickon Formation in Delaware (Plank et al., 2000; Schenck et al., 2000) with the type-Wissahickon in Philadelphia (Blackmer and Bosbyshell, 2002).

Zircons from the Wissahickon psammitic gneiss occur in a wide variety of morphologies, including spherical, oblong, and elongate shapes (Fig. 14A). The elongate grains are similar in morphology to igneous zircons; however, CL images reveal that these zircons are composed of rounded, oscillatory zoned cores that are truncated by mostly dark, unzoned overgrowths

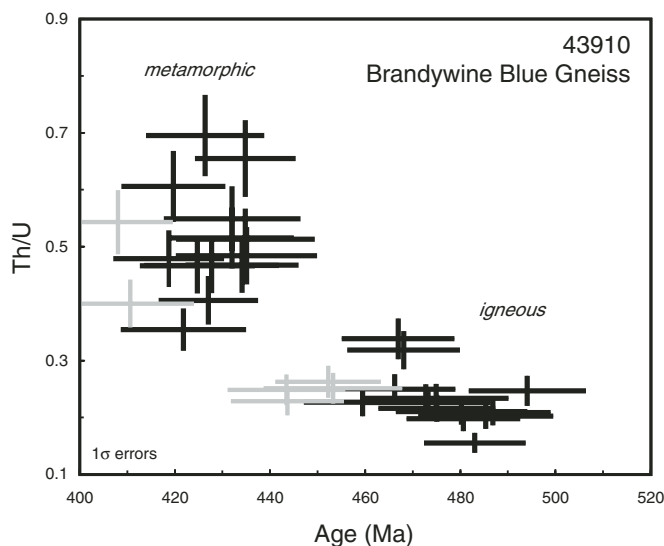


Figure 5. Th/U versus age of igneous and metamorphic zircons from the Brandywine Blue Gneiss. Gray error crosses from analyses were not used in age calculations (shown as dark-gray error ellipses in Fig. 4A). Uncertainty in Th/U is assumed to be $\pm 10\%$.

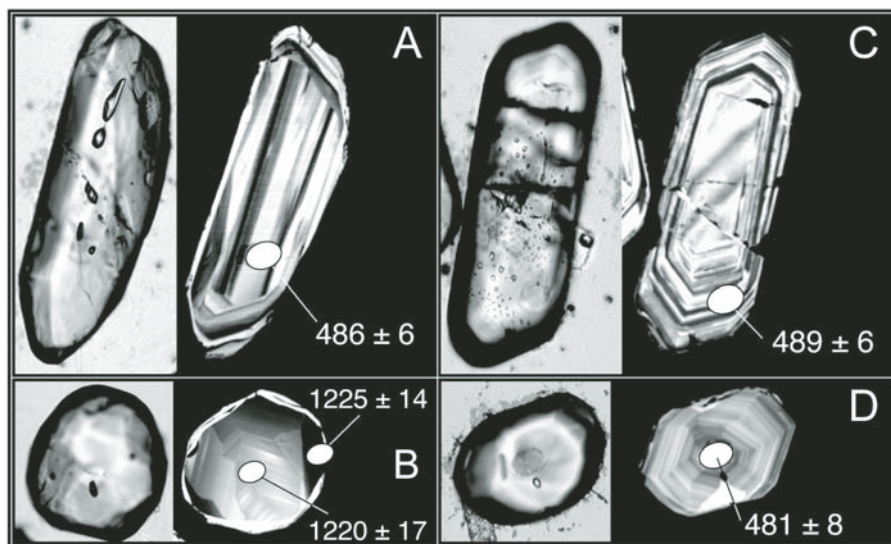


Figure 6. Transmitted-light (left side) and cathodoluminescence (right side) images of representative zircons from rocks of the Wilmington Complex. (A) Elongate igneous zircon showing zoning typical of volcanic origin, Windy Hills Gneiss (sample 43562). (B) Rounded detrital zircon containing truncated oscillatory zoning, Windy Hills Gneiss (sample 43562). (C) Elongate igneous zircon, Faulkland Gneiss (sample 43753). (D) Equant igneous zircon, Faulkland Gneiss (sample 43753).

(Fig. 14B). Although rounded zircons can form by sedimentary or magmatic processes, we conclude that all zircons from this sample of metasedimentary rock, including cores of elongate grains, are detrital in origin.

Five analyses are discordant (Fig. 15A) and are excluded from the relative probability diagram (calculated by Isoplot/Ex; Ludwig, 2003) showing the distribution of age populations (Fig. 15B). Concordant ages of zircons from sample 44069 range from ca. 900 to 1740 Ma (Table 2, Fig. 15B); most of the ages are between ca. 900 and 1400 Ma. Two major populations occur at ca. 950 and 1050 Ma, with subsidiary groups at ca. 1150 and 1275 Ma. This age distribution is similar to the ages of detrital zircons from the Windy Hills Gneiss (Fig. 15B); in both units the largest peak is at ca. 1050 Ma, and most of the other ages are between 0.9 and 1.4 Ga. The Windy Hills Gneiss also contains a few Neoproterozoic to early Paleozoic zircons that are younger than the youngest zircons found in the Wissahickon sample. Although the youngest detrital zircon in the Wissahickon (735 Ma) constrains the maximum age of deposition of the formation, better age constraints are provided by field evidence that show metasedimentary rocks of the Wissahickon to be interlayered with metavolcanic rocks of the Faulkland Gneiss, dated at 481 ± 4 Ma.

Monazite from the Wissahickon Formation (sample 44069) is fairly uniform in shape and color. Monazite grains are equant (discoidal) to subhedral and pale yellow. Most grains are completely unzoned in BSE, indicating chemical homogeneity. SHRIMP analysis of 13 grains results in an age of 426 ± 3 Ma. Two slightly older grains (analyses 44069m-1.1 and 44069m-7.1) and one younger age from an intermediate mantle zone (analysis 44069m-14.1) were excluded from the age calculation (Fig. 16A). TIMS data for monazite from sample 44069 are concordant. Nine concordant analyses yield a very precise Concordia Age (Ludwig, 1998, 2003) of 424.9 ± 0.4 (Table 3, Fig. 16B), in excellent agreement with the SHRIMP data. In contrast to monazite from the Faulkland Gneiss, no evidence was found for younger ages of monazite growth.

SUMMARY

Metaplutonic and metavolcanic rocks of the Wilmington Complex crystallized between ca. 475 and 485 Ma (Table 4). Detrital zircons in the Windy Hills Gneiss and Wissahickon Formation were derived primarily from Grenvillian crystalline basement, mostly of Mesoproterozoic age (1.0–1.4 Ga). These rocks were metamorphosed to amphibolite or granulite facies at ca. 425–

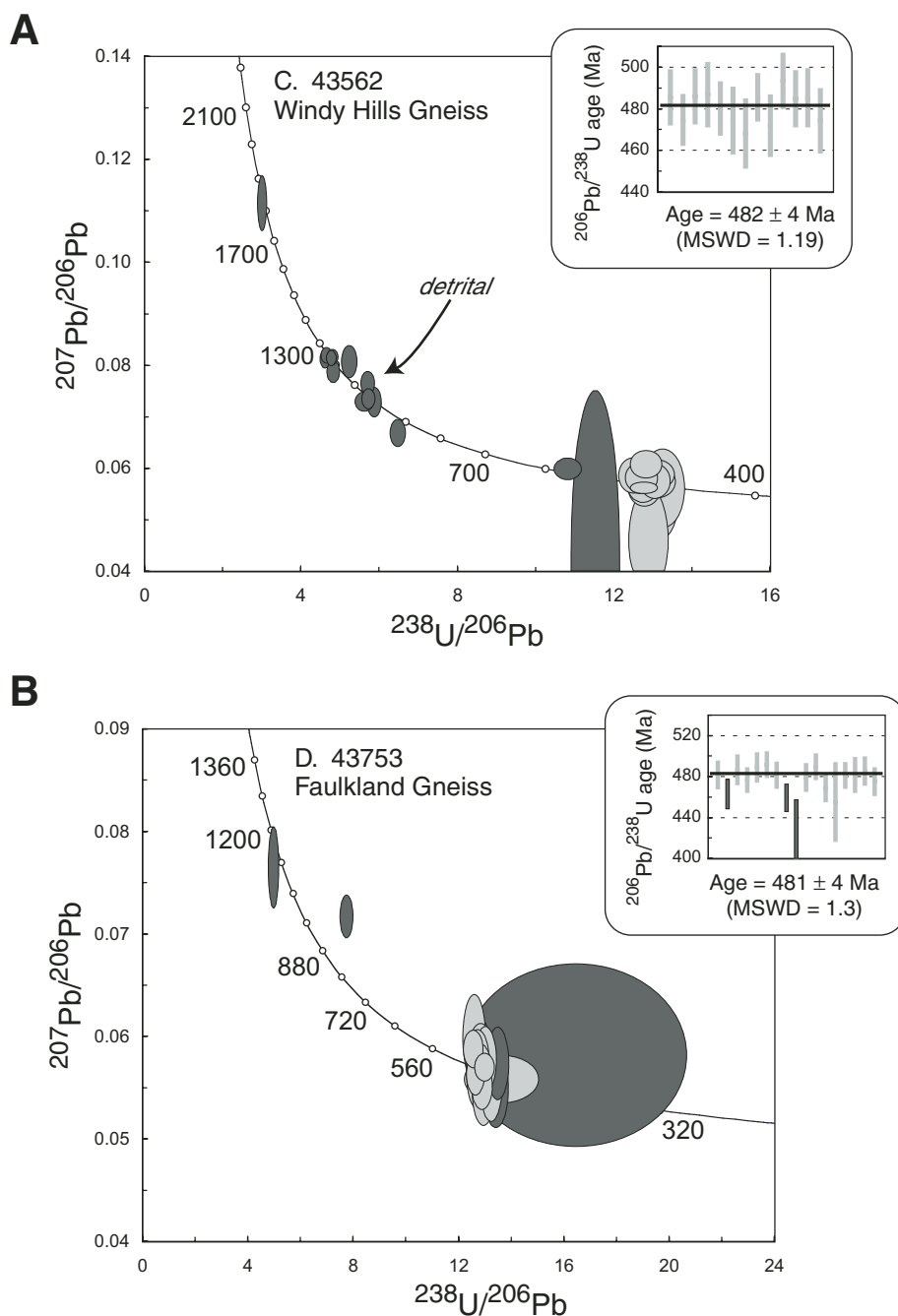


Figure 7. Tera-Wasserburg concordia diagrams and weighted average plots of zircon SHRIMP data. See Figure 4 caption for explanation of error ellipse shading. (A) Windy Hills Gneiss, Ordovician elongate igneous grains only. (B) Faulkland Gneiss.

435 Ma, synchronous with emplacement of the Arden Plutonic Supersuite. Isotopic data from the Faulkland Gneiss monazite suggest at least two younger episodes of monazite growth at ca. 410 and 400 Ma. Elongate and equant zircons formed during both igneous and metamorphic events. In addition, the Faulkland and Windy Hills Gneisses also contain equant detrital zircons.

DISCUSSION

Zircon Morphology

Igneous Zircon

Zircons from several units of the Wilmington Complex display a wide variety of morphologies. Euhedral, elongate zircons with oscillatory

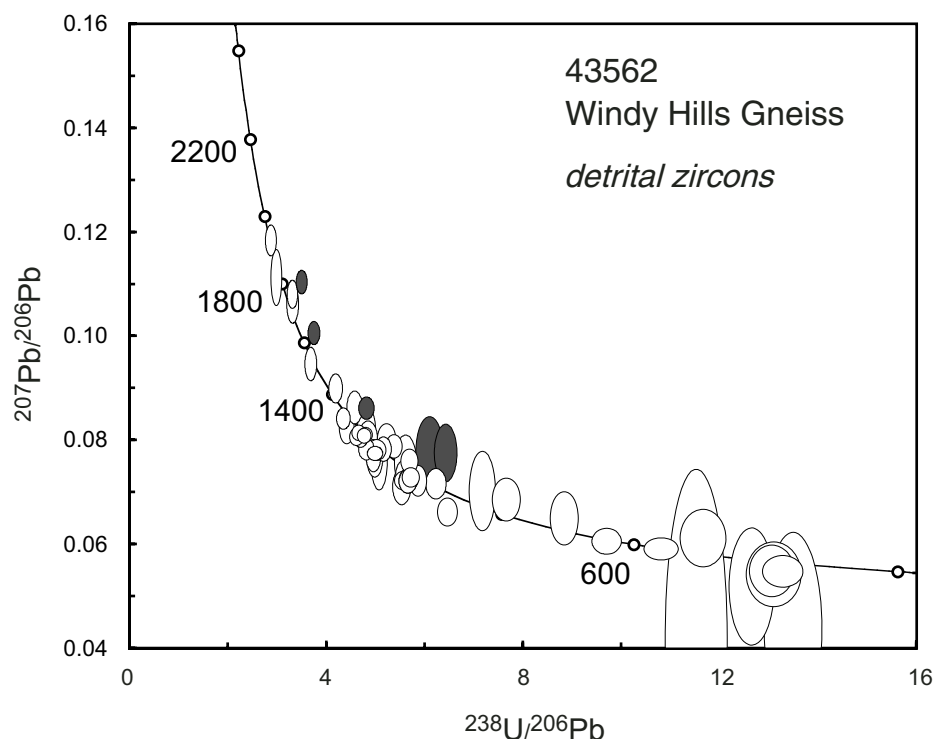


Figure 8. Tera-Wasserburg concordia diagram, Windy Hills Gneiss (sample 43562). Plot includes data from both analytical sessions. Discordant isotopic data, shown by gray error ellipses, were excluded from the relative probability plot (Fig. 15B).

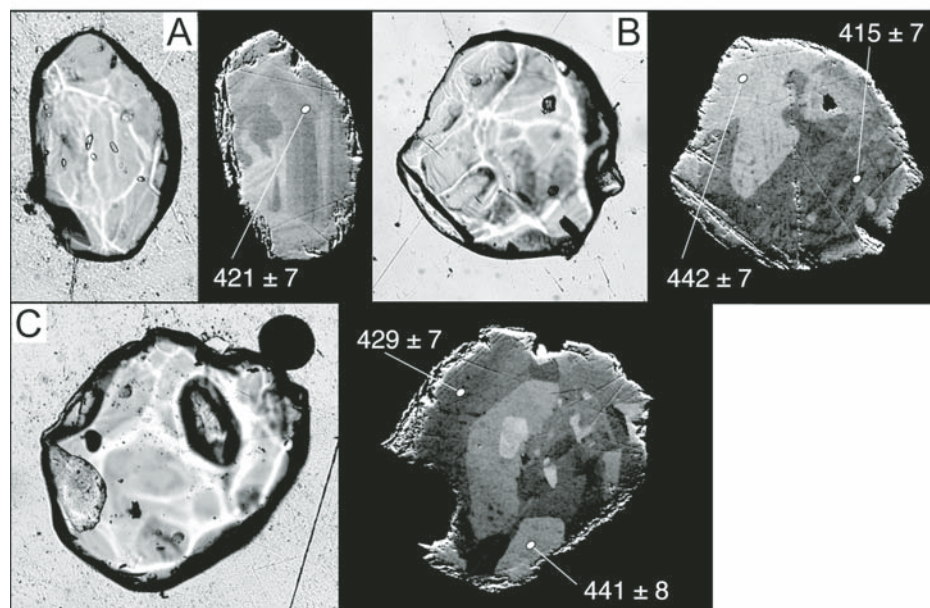


Figure 9. Transmitted-light (left side) and back-scattered electrons (right side) images of representative monazite from the Faulkland Gneiss (sample 43753). All ages were determined during session 1. The long axis of SHRIMP analytical spot is $\sim 15 \mu\text{m}$. (A) Mostly euhedral oscillatory zoned grain. (B) Unzoned, light-colored older area invaded by darker, patchwork-zoned overgrowth. (C) Patchwork-zoned, light-colored core area and darker rim.

zoning in CL, characteristic of igneous origin, were found in all dated samples of the Wilmington Complex (Table 4). In addition to elongate zircons, equant zircons of igneous origin are present in the Brandywine Blue Gneiss and Faulkland Gneiss. As shown by continuous, concentric, oscillatory zoning in CL, these are not simply broken zircons; they are euhedral and grew in this form in the magma. Among the many causes of morphologic variability in zircon (summarized in Speer, 1980, and shown graphically by Pupin, 1980), the most likely explanation for the occurrence of both equant and elongate zircon in the Brandywine Blue Gneiss may be local chemical variation in the magma. Alternatively, the growth rate of pyramidal faces may be retarded by adsorption of trace elements on those faces, thereby favoring formation of elongate prismatic grains (Vavra, 1990). Trace element analyses of the different zircon morphologies are required to further explore the potential causes of shape variations in igneous zircons found within samples of the Wilmington Complex.

Metamorphic Zircon

Metamorphic zircon usually forms during the prograde part of high-grade metamorphism and, if relatively pure (i.e., stoichiometric with low trace element concentrations and minimal metamictization), probably will not be reset even at granulite facies (cf. Hoskin and Black, 2000; Hoskin and Schaltegger, 2003). Because the closure temperature for diffusion of Pb in zircon is $>800^\circ\text{C}$ (Cherniak and Watson, 2001), U-Pb analyses of metamorphic zircons from amphibolite- and granulite-grade rocks will record the time of growth of the zircon during prograde metamorphism. This is in contrast to other isotopic systems, such as $^{40}\text{Ar}/^{39}\text{Ar}$ or Rb-Sr mineral ages, which represent times of cooling after peak high-grade metamorphic conditions at a particular closure temperature (in the range of ~ 300 – 500°C).

On the basis of age, we have concluded that two populations of zircons with very different morphologic characteristics formed in response to the same amphibolite-to-granulite facies metamorphic event. Elongate, prismatic metamorphic zircons (similar in appearance and zoning to igneous zircon) formed in the Brandywine Blue Gneiss, whereas metamorphic zircons in the Rockford Park Gneiss are equant and discoidal. This dichotomy in appearance probably is due to the composition of the host rock and may be indicative of formation at slightly different times and pressure-temperature conditions during metamorphism.

The Rockford Park Gneiss is a layered sequence of mafic and felsic gneisses. The dated

sample is from a relatively felsic layer (now composed primarily of quartz-plagioclase-orthopyroxene). The occurrence of typical metamorphic zircons (with patchwork zoning in CL) within this felsic layer suggests that they formed by a subsolidus reaction during prograde, amphibolite facies metamorphism (cf. Schaltegger et al., 1999). The source of Zr is unknown, although it may have been contained in igneous pyroxene in subjacent mafic layers.

We suggest a different process for the formation of metamorphic zircon in the Brandywine Blue Gneiss. This rock is a homogeneous felsic gneiss, now composed primarily of quartz and plagioclase, \pm minor orthopyroxene, clinopyroxene, and amphibole. Because of the high-grade metamorphic overprinting, the source of zircon in the Ordovician protolith is undetermined. On the basis of concentric, oscillatory zoning in Silurian zircon, these metamorphic grains probably formed only when granulite-grade metamorphic conditions (800 ± 50 °C; Srogi, 1988) were attained and local partial melting was initiated (Srogi et al., 1993). We suggest that igneous zircon was resorbed (perhaps as shown in Fig. 3A) and reprecipitated during anatexis (cf. Pan, 1997; Rubatto et al., 2001). Alternatively, other minor phases, such as apatite or xenotime, may have been involved in the formation of metamorphic zircon (Villaseca et al., 2003). During this anatectic process, Th and U were fractionated, as shown by different Th/U in igneous (0.15–0.35) and metamorphic (0.35–0.70) zircons.

Although metamorphic zircon characteristically has very low Th/U (i.e., <0.1 ; Hoskin and Schaltegger, 2003, and references therein), metamorphic zircons from units in the Wilmington Complex are atypical in their Th/U. Zircons interpreted as metamorphic in origin (on the basis of equant morphology, patchy zoning, and young age) in the Rockford Park Gneiss have about the same range of Th/U (mostly 0.6–0.9) as coexisting igneous zircon. Zircons from the Brandywine Blue Gneiss interpreted as metamorphic in origin (solely on the basis of young age) have Th/U that is greater than coexisting igneous grains (Fig. 5). We suspect that these metamorphic zircons retained igneous-like Th/U due to the absence of a Th-bearing phase (such as monazite, a common metamorphic mineral, which was not found in the mineral separations for either the Brandywine Blue Gneiss or Rockford Park Gneiss).

In summary, metamorphic zircons from the Rockford Park Gneiss and Brandywine Blue Gneiss formed in response to the same metamorphic event, but by very different processes and possibly at slightly different times that are not resolvable by SHRIMP analysis.

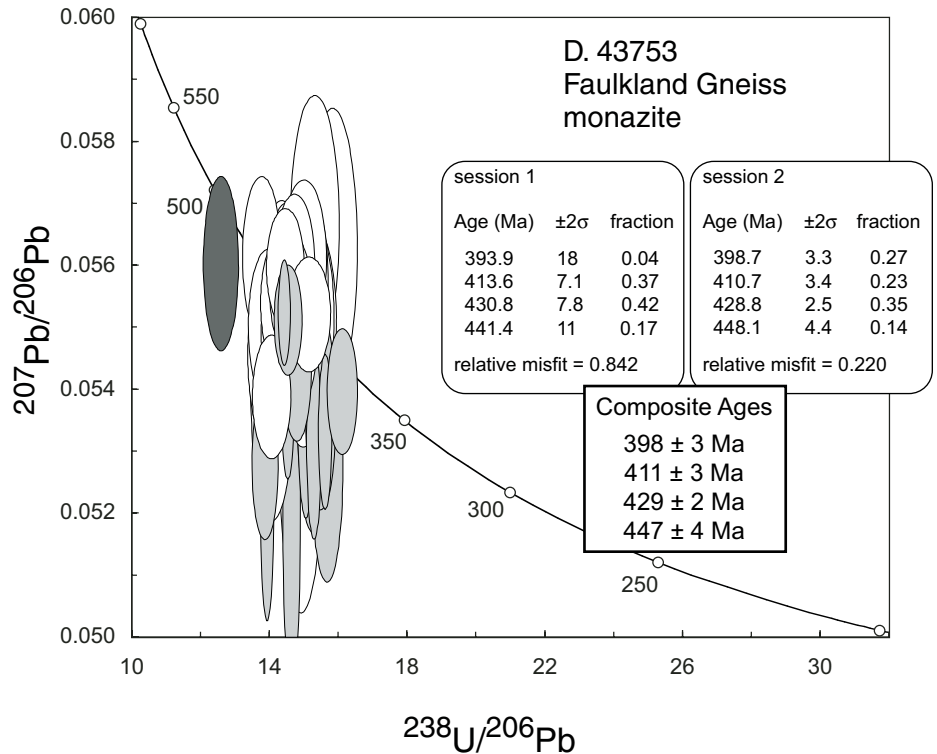


Figure 10. Tera-Wasserburg concordia diagram and weighted average plot of SHRIMP data for monazite from the Faulkland Gneiss. White ellipses are from session 1; light-gray ellipses are from session 2. Dark-gray ellipse was excluded from unmixing calculation.

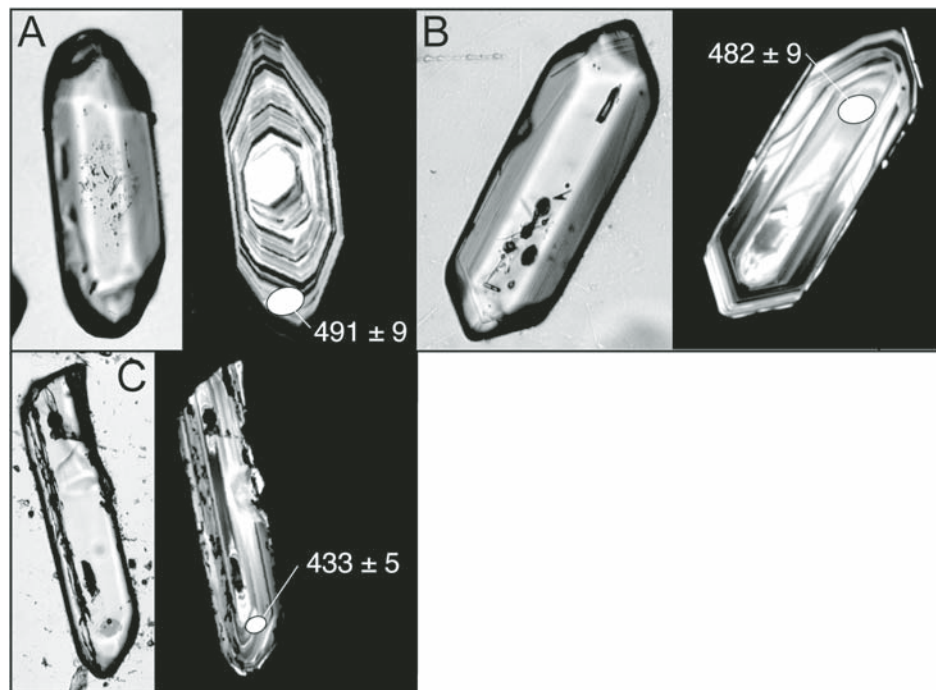


Figure 11. Transmitted-light (left side) and cathodoluminescence (right side) images of representative zircons from rocks of the Wilmington Complex. (A) Christianstead Gneiss (sample 42001). (B) Barley Mill Gneiss (sample 43817). (C) Ardentown Granitic Suite (sample 43959).

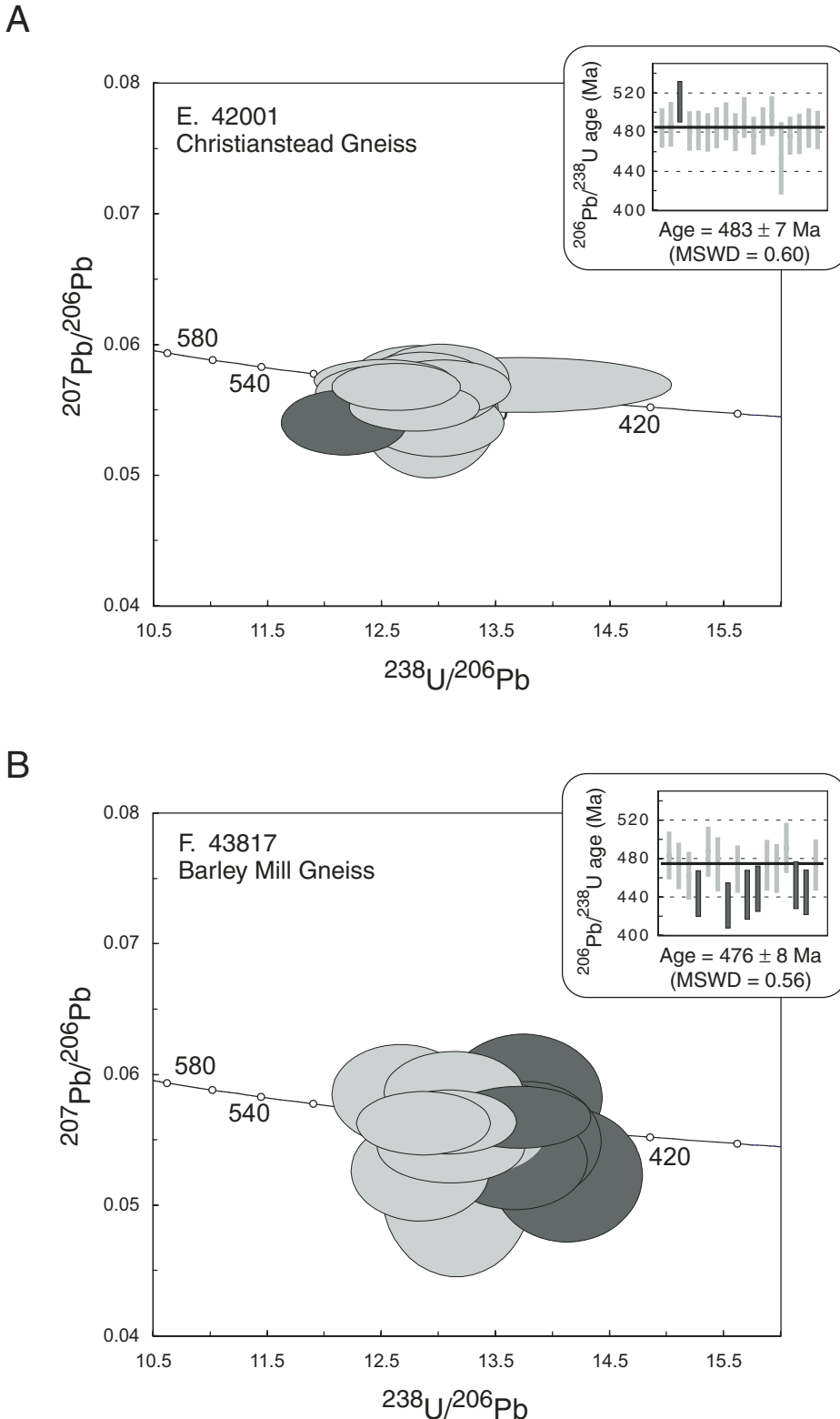


Figure 12. Tera-Wasserburg concordia diagrams and weighted average plots of zircon SHRIMP data. See Figure 4 caption for explanation of error ellipse shading. (A) Christianstead Gneiss. (B) Barley Mill Gneiss.

The Rockford Park Gneiss grew metamorphic zircon during prograde, amphibolite facies metamorphism, whereas the formation of metamorphic zircon in the Brandywine Blue Gneiss occurred at a higher metamorphic grade during partial melting. A similar scenario (subsolvus metamorphic crystallization followed by igneous growth during partial melting) is described for rocks in the Variscan Belt of eastern France (Schaltegger et al., 1999).

Timing of Metamorphism

The ages of metamorphic zircons in rocks of the Wilmington Complex (432 ± 6 and 428 ± 4 Ma) indicate the time of prograde amphibolite to granulite metamorphism of the Rockford Park Gneiss and Brandywine Blue Gneiss, respectively, and by inference the other units of the complex. This Silurian metamorphic event also is recorded by ages of monazite from the Faulkland Gneiss (429 ± 2 Ma) and Wissahickon Formation (426 ± 3 Ma, 424.9 ± 0.5 Ma). At about the same time (435 ± 5 Ma), rocks of the Arden Plutonic Supersuite were intruded. It is doubtful that the metamorphism of the Wilmington Complex is a contact effect related to emplacement of the Arden. More likely, regional-scale tectonism probably caused crustal melting at depth, leading to emplacement of the Arden plutonic rocks and the metamorphism of the Wilmington Complex. The nearby Springfield Granite has a similar age of about 427 Ma (J. Aleinikoff, 2004, personal commun.).

Zircon and monazite also have an older age in common. SHRIMP analyses of zircon from four Ordovician rocks of the Wilmington Complex (samples 43910 [Brandywine Blue Gneiss], 43909-2 [Rockford Park Gneiss], 43753 [Faulkland Gneiss], and 43817 [Barley Mill Gneiss]) yielded a few intermediate ages between the accepted emplacement age of 475–485 Ma and the well-documented Silurian metamorphic event at ca. 425–435 Ma (see Figs. 4, 7, and 12). Although we are unable to ascribe a specific origin to these intermediate ages, it is interesting that they are about the same age as the oldest monazite age (447 ± 4 Ma) from the Faulkland Gneiss, suggesting that this may be a period of growth. It is also possible that the intermediate ages are due to Pb loss in Ordovician zircon and are coincidentally similar to the 447 ± 4 Ma monazite age.

All of the ages recorded in monazite from the Faulkland Gneiss correspond to known events in the eastern United States. The oldest age is typical of Late Ordovician arc rocks in the Bronson Hill terrane of New England (Tucker and Robinson, 1990) and locally is expressed as the emplacement age of tonalite of the Norbeck

Intrusive Suite in Maryland (449 ± 7 Ma; Aleinikoff et al., 2002). The ages of 411 ± 3 Ma and 399 ± 3 Ma, which might also be recorded in analyses of zircons from the Brandywine Blue Gneiss (sample 43910), Faulkland Gneiss (sample 43753), and pyroxene quartz diorite of the Ardentown Granitic Suite (Figs. 4, 7, and 12), may be related to regional deformation associated with the Acadian orogeny. Despite numerous SHRIMP analyses of zones, including cores and overgrowths in individual monazite grains, no younger ages were found, in contrast to electron microprobe analysis (EMPA) data suggesting an age of ca. 360–365 Ma for some monazite overgrowths from the Wissahickon Formation in Philadelphia (Bosbyshell et al., 2001; Crawford et al., 2001). The EMPA monazite ages and the $^{40}\text{Ar}/^{39}\text{Ar}$ biotite and Rb-Sr biotite data suggest that the Late Devonian event was the time of final heating and subsequent cooling of the rocks of the Wilmington Complex.

Provenances of Detrital Zircons in the Windy Hills Gneiss and Wissahickon Formation

Most of the dated detrital zircons from the Windy Hills Gneiss and Wissahickon Formation have ages in the range of 0.90–1.4 Ga. These ages are typical of crystalline rocks of the Grenville province that occur in a belt of basement inliers in the eastern United States from Vermont to Georgia. According to Plank et al. (2000), the stratigraphically lower part of the Wissahickon Formation, from which our sample 44069 was collected, was deposited in an ocean basin offshore from Grenville crystalline rocks of the Mill Creek Nappe and Avondale and West Chester massifs located just northwest of the Wilmington Complex. Concordia plots of discordant TIMS data from multigrain fractions of zircons from basement gneisses of the West Chester and Avondale massifs suggest ages of ca. 0.98–1.2 Ga (Grauert et al., 1973, 1974). The largest age population (1.05 Ga) of detrital zircons from the Wissahickon Formation and Windy Hills Gneiss is also similar to ages of basement rocks in the northern and central Blue Ridge, Virginia (Aleinikoff et al., 2000; Tollo et al., 2004), Reading Prong and Hudson Highlands, eastern Pennsylvania to southeastern New York (Walsh and Aleinikoff, 2004; Ratcliffe and Aleinikoff, 2001), and the Adirondacks, northern New York (McLelland et al., 2001). Rocks of comparable age also occur in basement uplifts of the southern Appalachians (Carrigan et al., 2003).

On the basis of U-Pb ages of detrital zircons in the Wissahickon Formation, Laurentian crustal rocks are the most likely sources of

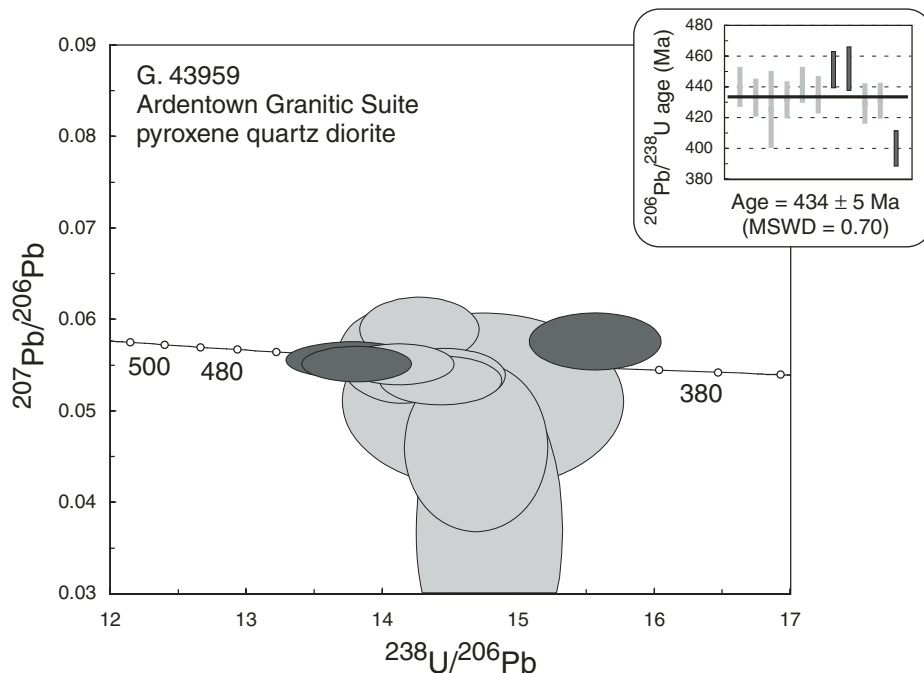


Figure 13. Tera-Wasserburg concordia diagram and weighted average plot of zircon SHRIMP data from the Ardentown Granitic Suite. See Figure 4 caption for explanation of error ellipse shading.

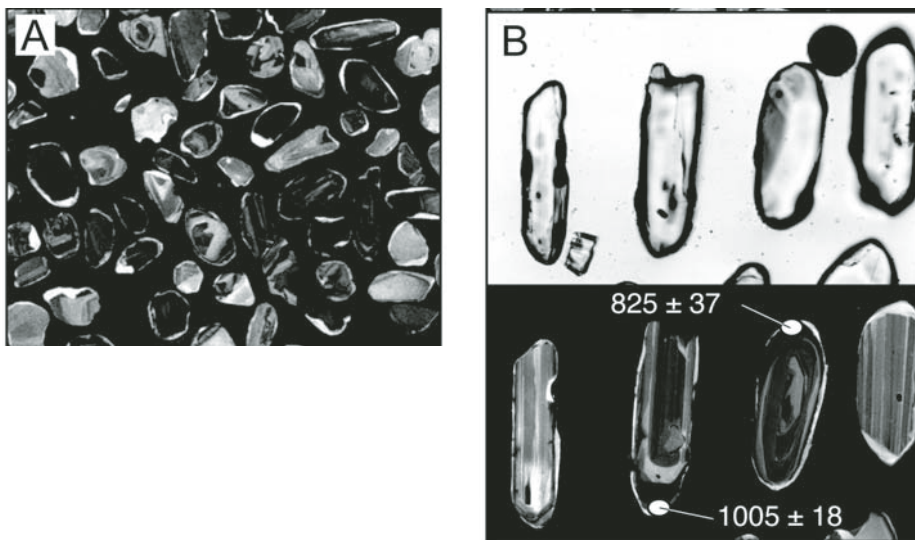


Figure 14. Detrital zircons from the Wissahickon Formation (sample 44069). (A) Cathodoluminescence (CL) image of random selection of detrital grains, showing rounded, irregular shapes and truncated zoning. (B) Transmitted light and CL images of elongate zircons. All ages, including those of metamorphic overgrowths, are Proterozoic.

sediment deposited in an accretionary prism near the magmatic arc that ultimately became the Wilmington Complex. Thus, in the Early Ordovician, this arc may have been situated near the Laurentian margin.

Tectonic History of the Wilmington Complex Island Arc

The protoliths of metavolcanic and metaplutonic rocks within the Wilmington Complex

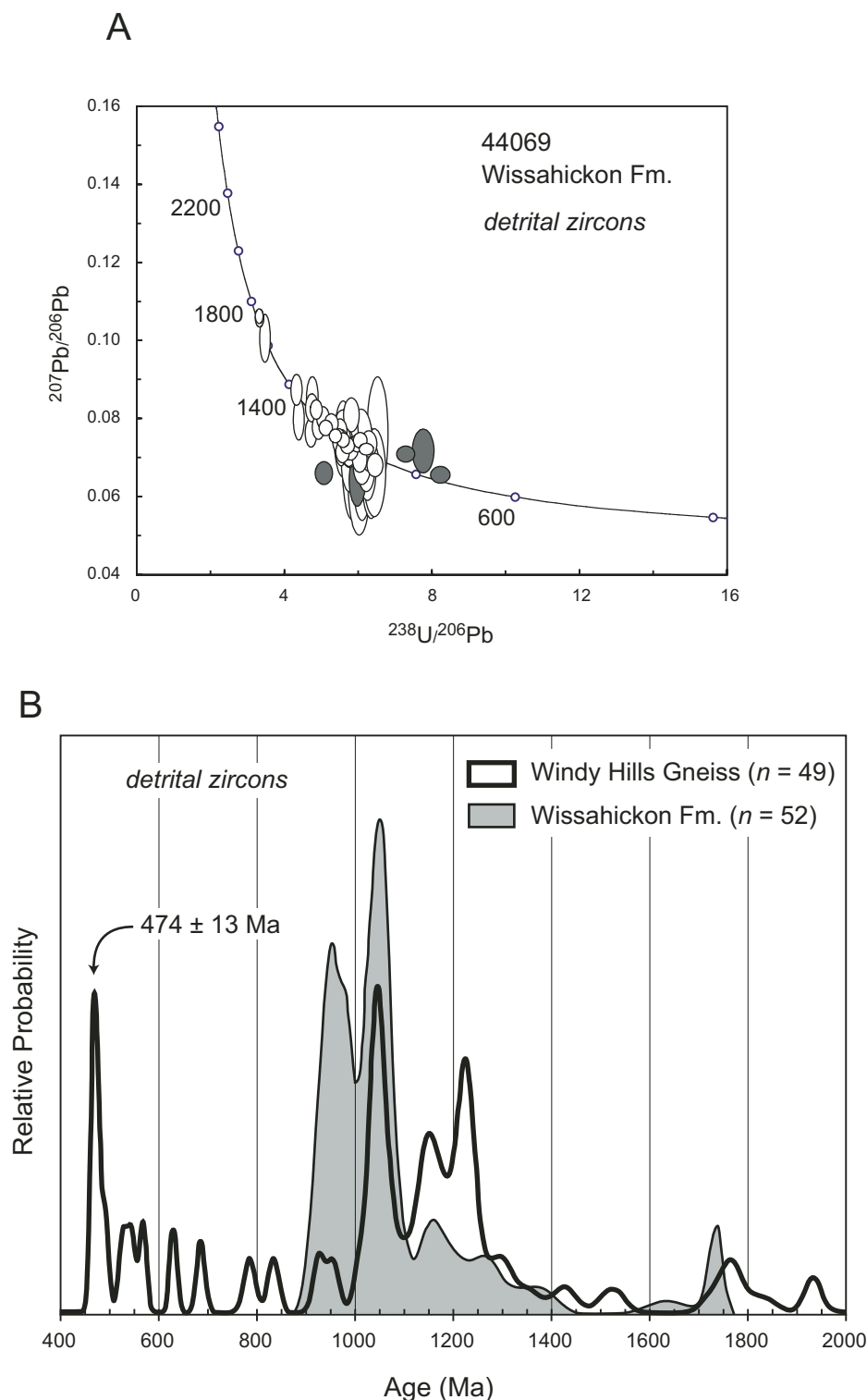


Figure 15. Plots of U-Pb data from detrital zircons. (A) Tera-Wasserburg concordia diagram, Wissahickon Formation (sample 44069). Discordant isotopic data, shown by gray error ellipses, were excluded from the relative probability plot (Fig. 15B). (B) Combined relative probability plot of detrital zircons from the Windy Hills Gneiss and Wissahickon Formation.

were emplaced between ca. 470 and 485 Ma. The close geographic and geochronologic association of all members of the Wilmington Complex suggests a cogenetic origin. Despite subsequent high-grade metamorphism, geochemical fingerprinting of mafic layers (now mafic gneisses and amphibolites) within members of the Wilmington Complex indicates that all of these rocks probably formed at a convergent margin as a magmatic arc (Plank et al., 2001). On the basis of spider diagrams (Rollinson, 1993) and various discrimination plots utilizing concentrations of immobile elements (Pearce and Cann, 1973; Shervais, 1982; Mullen, 1983), these mafic rocks are subdivided into several geochemical groups, including boninitic basalt, low-K arc tholeiite, and backarc basin basalt.

Interpretations of Srogi (1988), Wagner and Srogi (1987), Plank et al. (2000), and Bosbyshell et al. (2001), combined with the new geochronologic data, suggest that (1) sediments (now the Wissahickon Formation) were deposited in forearc and (or) accretionary wedge settings, (2) this sedimentation was at least partly contemporaneous with formation of the magmatic arc, including volcanic rocks now exposed as Rockford Park Gneiss, Windy Hills Gneiss, and Faulkland Gneiss, and plutonic rocks now exposed as the Brandywine Blue Gneiss, Barley Mill Gneiss, and Christianstead Gneiss, and (3) these events occurred in the Early Ordovician. Intrusion of mafic dikes of boninitic affinity (thought to belong to the Rockford Park Gneiss sequence) into rocks mapped as the Wissahickon Formation suggests that the deposition of at least some of the Wissahickon preceded extrusion of the Wilmington Complex volcanic rocks. Burial of the Wilmington Complex culminated in the formation of metamorphic mineral assemblages indicative of depths of 15–20 km and temperatures of 700–850 °C (amphibolite to granulite facies metamorphism) (Wagner and Srogi, 1987; Srogi, 1988; Srogi et al., 1993). A previous model of overthrusting of the Wilmington Complex onto the Wissahickon Formation (Wagner and Srogi, 1987) now is considered unlikely (Bosbyshell et al., 1999), because these units are thought to have been juxtaposed fairly early in the history of the arc. The currently preferred model for the cause of metamorphism is crustal thickening due to collision of the arc with Laurentia. The occurrence of Grenville-age detrital zircons in the Windy Hills Gneiss and Wissahickon Formation suggests proximity of the arc to the Laurentian margin in the Ordovician. However, perhaps final collision (and crustal thickening) only occurred in the Silurian, ~50 m.y. after formation of the arc. The Arden Plutonic Supersuite, composed of contemporaneous mantle-derived mafic

rocks and associated crustal-derived granitic rocks, plus similar plutons, was emplaced at ca. 425–435 Ma, perhaps due to slab breakoff and rollback or backarc rifting of a new subduction zone to the east. Heating and/or metamorphism of the deeply buried Ordovician arc rocks also occurred at this time.

Comparison with Other Arcs

The presence of a Mesoproterozoic detrital zircon population within the sample of Windy Hills Gneiss suggests that the arc volcanic rocks of the Wilmington Complex did not form as an isolated tectonic element but rather had a sedimentary link to a continental mass. Ages of ca. 1.0–1.3 Ga are typical of Grenville basement rocks now exposed in a number of inliers occurring in a northeast-southwest belt from Georgia to Vermont, which includes the Mill Creek Nappe and West Chester and Avondale massifs immediately to the northwest of the Wilmington Complex (Grauert et al., 1973, 1974) and the Baltimore Gneiss domes to the southwest (Aleinikoff et al., 1997, 2004). The detrital zircon age data suggest that arc rocks of the Wilmington Complex may have formed as part of an Early Ordovician peri-Laurentian arc (Mac Niocaill et al., 1997). Paleogeographic reconstructions (cf. McKerrow et al., 2000, and references therein) show that this arc extended through what is now Massachusetts across eastern Canada, Ireland, and Scotland and into east Greenland. In the United States, rocks of similar Early Ordovician age and with Mesoproterozoic inheritance or detrital component include the Shelburne Falls arc in Massachusetts and Vermont (Karabinos et al., 1998), the Chopawamsic and Milton terranes of Virginia (Coler et al., 2000), and tonalitic rocks of the Maryland and Virginia Piedmont (Aleinikoff et al., 2002). Sedimentologic and geochronologic data in Georgia and Alabama suggest the arc may have extended as far south as the present-day southern Appalachians, although a lack of mafic arc-related volcanic rocks in North Carolina implies either that evidence of subduction zone magmatism is buried or eroded, or that the arc was discontinuous (Thomas et al., 2001). It formed at about the same time as Early Ordovician arc rocks of the Exploits terrane of Newfoundland (cf. MacLachlan and Dunning, 1998) and the peri-Avalonian arc (summarized in Mac Niocaill et al., 1997), but presumably in a more southerly location in the Iapetus Ocean.

Karabinos et al. (1998) proposed that the Taconic orogeny commenced in New England during the collision of the Shelburne Falls arc with Laurentia at ca. 465–470 Ma. This collision ultimately resulted in formation of a new

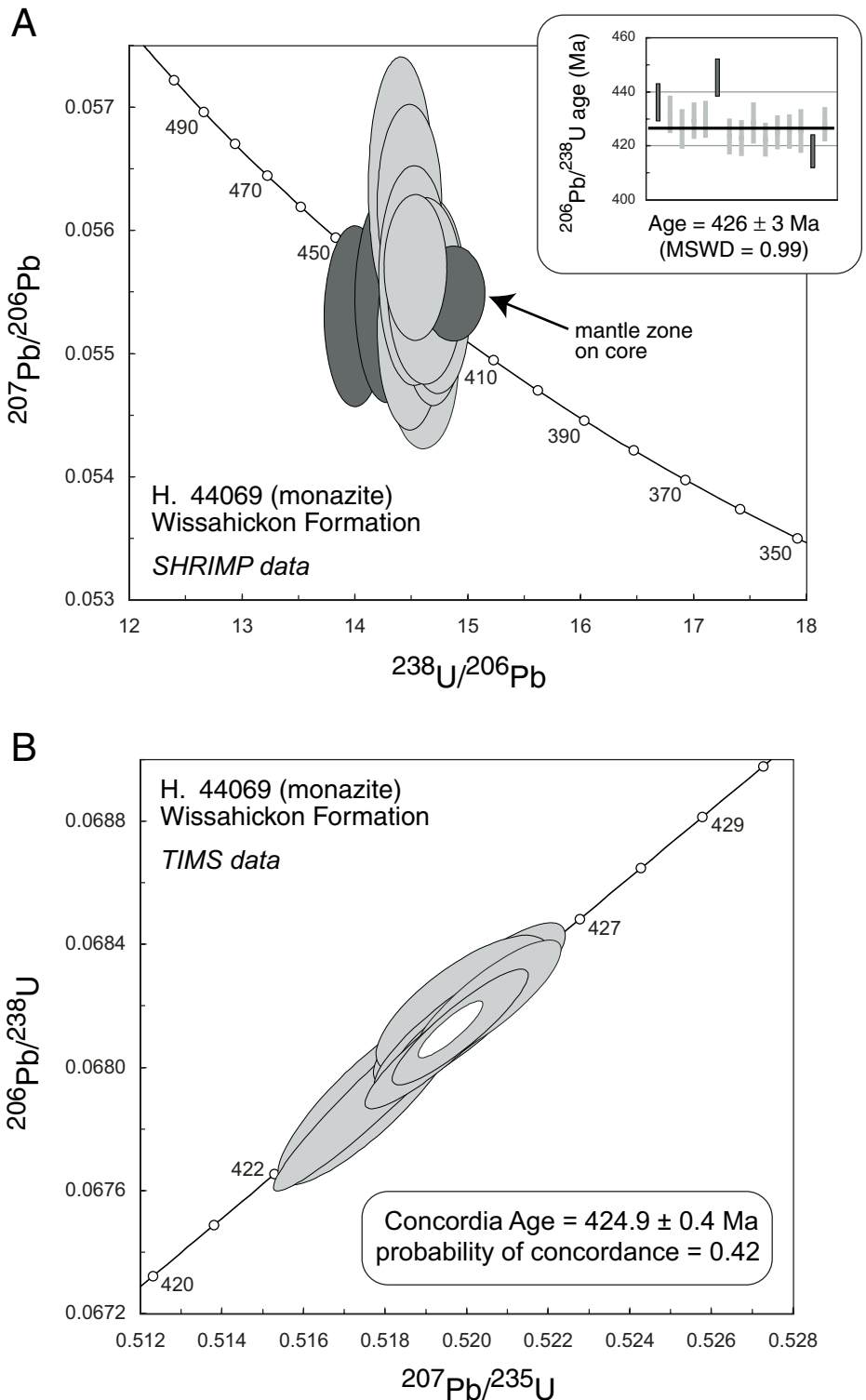


Figure 16. Concordia plots of U-Pb isotopic data for monazite from the Wissahickon Formation (sample 44069). (A) Tera-Wasserburg plot of SHRIMP data. (B) Conventional plot of TIMS data. White ellipse represents Concordia Age and uncertainty, based on calculation using concordant data (Ludwig, 2003).

island arc (Bronson Hill arc), with a polarity reversal (west-dipping subduction zone) ca. 445–450 Ma, now exposed in central Massachusetts and western New Hampshire (cf. Zartman and Leo, 1985; Tucker and Robinson, 1990; Moench and Aleinikoff, 2002). A similar tectonic scenario, at a somewhat younger age, has been proposed for rocks of the Wilmington Complex (Bosbyshell et al., 2001). Our geochronologic data do not provide constraints for the timing of arc-continent collision; however, the age of formation of metamorphic zircons in the Rockford Park Gneiss indicates that the Wilmington Complex underwent amphibolite to granulite facies metamorphism beginning at 432 ± 6 Ma.

Possible central Appalachian counterparts for the Bronson Hill arc may be volcanic units within the James Run Formation (459 ± 5 ; 465 ± 5 Ma), Relay Felsite (461 ± 5 Ma), Chopawamsic Formation (454 ± 5 Ma), and Quantico Formation (451 ± 6 Ma) (Horton et al., 1998) to the south of the current study. All of these volcanic rocks are distinctly younger than arc rocks of the Wilmington Complex.

CONCLUSIONS

1. Metaigneous rocks of the Wilmington Complex (of plutonic and volcanic origin) crystallized at ca. 475–485 Ma. These rocks contain zircons that have a wide variety of external morphologies and internal zoning patterns.
2. Amphibolite to granulite facies metamorphism of rocks of the Wilmington Complex occurred at ca. 425–435 Ma, at about the same time as emplacement of mantle-derived mafic rocks of the Arden Plutonic Supersuite.
3. Elongate and equant zircons in rocks of the Wilmington Complex formed by both igneous and metamorphic processes. Cathodoluminescence zoning is useful in distinguishing zircon origin, especially when used in the context of other petrologic and petrographic data. Similar metamorphic ages are recorded by elongate zircons (with concentric oscillatory zoning) that formed during local anatexis of felsic gneiss and by equant zircons (with patchwork zoning) in a felsic gneiss member of a layered gneiss sequence metamorphosed to granulite grade.
4. Monazite growth occurred during at least four stages (447 ± 4 , 429 ± 2 , 411 ± 3 , and 398 ± 3 Ma). Each of these episodes can be related to regional events in the central Appalachians.
5. On the basis of similar Early Ordovician ages and apparent tectonic setting, island-arc rocks of the Wilmington Complex are considered to be part of a belt of peri-Laurentian arc rocks that stretches from at least central Virginia (Chopawamsic and Milton terranes) to Vermont

(Shelburne Falls arc), and possibly as far north-east (present-day coordinates) as Newfoundland (Exploits terrane).

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