



# Paleozoic sediment dispersal before and during the collision between Laurentia and Gondwana in the Fort Worth Basin, USA

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## ABSTRACT

**We report detrital zircon U-Pb ages in the Fort Worth Basin (southern USA) aimed at understanding sediment dispersal patterns on the southern margin of Laurentia before and during the Laurentia-Gondwana collision. The ages from two Cambrian fluvial-marginal marine sandstone and six Pennsylvanian deltaic-fluvial sandstone samples span from Archean to early Paleozoic time. In the Cambrian sandstones, 80% of zircons are of Mesoproterozoic age (1.451–1.325 Ga) and 18% are of Grenvillian age. The high abundance of the Mesoproterozoic population suggests that the grains were dispersed by a local river draining the midcontinent granite-rhyolite province located in the Texas Arch to the northwest of the Fort Worth Basin. In the Pennsylvanian sandstones, 26% of zircons are of Archean–early Mesoproterozoic age, 47% are of Grenvillian age, 15% are of Neoproterozoic–earliest Paleozoic age (800–500 Ma), and 10% are of early Paleozoic age (500–318 Ma), indicating a different dispersal pattern during the Pennsylvanian relative to the Cambrian. Compared to other early Paleozoic detrital zircon records on the southern margin of Laurentia, our Pennsylvanian sandstones have a distinct age peak at ca. 650–550 Ma, which we interpreted to be a result of transport by local rivers draining a peri-Gondwana terrane, most likely the Sabine terrane in the Ouachita orogen. The high abundance of Grenvillian zircons reflects either direct transport from the Appalachians by an axial river or recycling from Mississippian–Pennsylvanian sedimentary rocks incorporated in the Ouachita orogenic front. The similarity of detrital zircon age distributions in the Fort Worth Basin, the Arkoma Basin, and the southern Appalachian forelands seems to favor sediment dispersal by a major river with headwaters in the southern Appalachians.**

## INTRODUCTION

Subduction of the southern Laurentian margin underneath Gondwana and the subsequent collision between the two continents caused the Alleghanian-Ouachita-Marathon orogenies and flexural subsidence of several peripheral foreland basins in southern Laurentia during the late Paleozoic (e.g., Walper, 1982; Erlich and Coleman, 2005; Thomas, 2006; Elebiju et al., 2010).

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The contractional tectonics may have accumulated enough intracontinental stress to reactivate paleo-transform faults in western Laurasia, leading to the basement-involved Ancestral Rocky Mountain (ARM) orogeny (Dickinson, 2000; Dickinson and Lawton, 2003). Sediment dispersal patterns and paleogeographic reconstruction of the southern margin of Laurentia are fundamental to the understanding of the tectonic configuration before and during the collision, as well as the intracontinental responses to the collision.

In north-central Texas, USA, the Fort Worth Basin evolved from a passive margin to a foreland basin during the late Paleozoic (Thomas, 1977; Walper, 1982). Numerous studies have been conducted on the Paleozoic basin fill in the Fort Worth Basin in order to understand petroleum systems of the region (e.g., Morris, 1974; Pollastro et al., 2003, 2007; Hentz et al., 2012), yet the provenances of the basin fill and the paleogeography before and during the collision remain poorly constrained. Two hypotheses have been proposed to explain the provenance of the Paleozoic basin fill in the Fort Worth Basin. Graham et al. (1975) proposed that the Paleozoic highland of the Appalachians funneled sediments into its foreland, and the sediments were then transported by a large axial river and dispersed into remnant ocean basins. This idea was further supported by neodymium isotope studies in the Arkoma Basin in Oklahoma and Arkansas, and Marathon Basin in west Texas (Gleason et al., 1994, 1995); detrital zircon U-Pb ages in the Marathon foreland in west Texas (Gleason et al., 2007); and <sup>40</sup>Ar/<sup>39</sup>Ar analysis of detrital muscovite in the Black Warrior Basin in Alabama, USA (Uddin et al., 2016). As summarized in Archer and Greb (1995), the most direct physical evidence for the river system is Early Pennsylvanian paleovalleys with 20–70 m of incision and south-southwestward paleoflow directions in the central Appalachian Basin in Kentucky and Indiana. Other studies have suggested that the large river originated in the Appalachians, instead of flowing through the Appalachian-Ouachita foreland, and drained through or around the Illinois Basin before entering the Ouachita front (Thomas, 1997; Sharrah, 2006; Xie et al., 2016). However, based on fractional volume of sand and interpretation of westward and southwestward propagation of deltas, the siliciclastic detritus in the Fort Worth Basin have been suggested to be from local sources, including the basin-bounding Ouachita orogen to the east and the basement-involved ARM uplifts to the north of the basin (Lovick et al., 1982; Walper, 1982; Hentz et al., 2012).

This study determines sediment provenance of Cambrian and Pennsylvanian sandstones in the Fort Worth Basin in southern Laurentia using detrital

zircon U-Pb ages. By combining our data with published detrital zircon data in eastern and southern Laurentia, we infer dispersal patterns of siliciclastic sediments on the southern margin of Laurentia, and reconstruct the paleogeography before and during the suturing between Gondwana and Laurentia.

## ■ GEOLOGICAL SETTING

The Fort Worth Basin is a foreland basin formed on the southern Laurentian continental shelf during the late Paleozoic Ouachita orogeny (e.g., Graham et al., 1975; Walper, 1977; Nelson et al., 1982; Thomas and Viele, 1983). The basin is bounded by the basement-involved Red River and Muenster uplifts to the north, the Ouachita thrust belt to the east, and the Llano uplift to the south, and it shallows gently toward the Bend Arch to the west (Fig. 1).

After the Neoproterozoic rifting of the supercontinent Rodinia, the southern margin of Laurentia gradually evolved into a passive margin of the Iapetus Ocean during the early Cambrian (Houseknecht and Matthews, 1985; Poole et al., 2005; Cawood and Nemchin, 2001). In association with the opening of the Iapetus Ocean, a rift zone was initiated in southern Oklahoma (Keller and Stephenson, 2007; Hanson et al., 2013) and formed the southern Oklahoma aulacogen in the vicinity of the Fort Worth Basin (Burke, 1977; Perry, 1989). The closure of the Iapetus Ocean started during the Early Ordovician, and was associated with the separation of several arc terranes (e.g., Avalonia, Ganderia, Carolina, Meguma, Suwannee) from the continental margin of northern Gondwana (e.g., Nance et al., 2002; Keppie et al., 2003; Murphy et al., 2006; Nance and Linnemann, 2008; Pollock et al., 2009; van Staal et al., 2009; Mueller et al., 2014). These terranes were later accreted to Laurentia during the closure of the Iapetus and Rheic Oceans and suturing between Gondwana and Laurentia as early as the Early Devonian. By the Late Devonian, Laurentia, Baltica, and Avalonia-Carolina had collided to form Laurussia. The suturing mechanism between Laurussia and Gondwana has long been debated. While most studies have agreed that the southern margin of Laurentia subducted underneath Gondwana (e.g., Thomas, 2004; Poole et al., 2005; Nance and Linnemann, 2008), other studies have suggested that the Laurentian plate was the upper plate before the collision (e.g., Domeier and Torsvik, 2014). The final closure of the Rheic Ocean resulted in the complete burial of the Laurentian platform by thick synorogenic sedimentation during the Pennsylvanian in the newly formed foreland basins (e.g., Hatcher, 1989; Viele and Thomas, 1989; Thomas, 2004).

The Alleghanian orogeny that occurred in southeastern Laurentia is the youngest orogenic event in the Appalachian orogen (Nance et al., 2010). The Ouachita orogeny occurred in southern Laurentia and represents the southwestward extension of the Alleghanian orogeny (Loomis et al., 1994; Poole et al., 2005), or is possibly coeval with the Alleghanian orogeny (Thomas, 1977; Robinson et al., 2012). The Ouachita orogen extends from Mississippi westward to southeastern Oklahoma, bends southward in eastern Texas, and is contiguous with the Marathon uplift in western Texas (Graham et al., 1975; Houseknecht and Matthews, 1985; Loomis et al., 1994; Poole et al., 2005)

(Fig. 1). Today, the mountains are mostly buried underneath Mesozoic and Cenozoic strata of the Gulf Coastal Plains (Houseknecht and Matthews, 1985; Loomis et al., 1994), and exposed only in the Marathon and Solitario uplifts in west Texas and in the Ouachita Mountains in Arkansas and Oklahoma (Thomas and Viele, 1983; Noble, 1993) (Fig. 1).

Contractional tectonics during the late Paleozoic also reactivated rift-related faults of the southern Oklahoma aulacogen and formed the northwest-striking Red River and Muenster uplifts as part of the Amarillo-Wichita uplift (Walper, 1982; Keller et al., 1989; Montgomery et al., 2005; Elebiju et al., 2010; Alsalem et al., 2017). Reactivation of early Paleozoic normal faults also caused the initial rise of the basement-involved Llano uplift in southern Laurentia (Erlich and Coleman, 2005). Exhumation of the Llano uplift continued into the Late Pennsylvanian, and may have tilted the strata in the Fort Worth Basin westward (Thomas, 2003). The Fort Worth Basin may have experienced exhumation during the late Permian–Jurassic as a result of the opening of the Gulf of Mexico (Jarvie et al., 2005; Ewing, 2006; Stern and Dickinson, 2010).

## ■ STRATIGRAPHY AND SAMPLES

The Cambrian–lower Permian sedimentary rocks in the Fort Worth Basin are up to ~3.7 km thick (Montgomery et al., 2005). In north-central Texas, the late Cambrian strata unconformably overlie Proterozoic gneiss and schist as well as granitic intrusions (Stenzel, 1935). The Paleozoic strata are unconformably overlain by Lower Cretaceous strata in the eastern part of the Fort Worth Basin (Fig. 1C). Based on depositional environments and tectonic histories, Montgomery et al. (2005) divided the Paleozoic strata in the Fort Worth Basin roughly into three intervals: (1) late Cambrian to Ordovician platform strata deposited in a passive continental margin; (2) Mississippian shallow marine rocks deposited during the exhumation of the Amarillo-Wichita uplift; and (3) Pennsylvanian–lowermost Permian interbedded shallow marine and deltaic deposition associated with the development of Ouachita orogen. The Paleozoic strata thin rapidly toward the southwest and crop out in the Llano uplift (Turner, 1957; Moore, 1959; Feray and Brooks, 1966; Erlich and Coleman, 2005; Montgomery et al., 2005; Alsalem et al., 2017). Below we briefly describe the clastic units that we sampled. The lithofacies of each sample and the sample locations are summarized in Table 1.

Two sandstone samples (S1 and S2) were collected from the Hickory Sandstone Member of the upper Cambrian Riley Formation in the south side of the Fort Worth Basin (Fig. 1C; Table 1). The member comprises fluvial–shallow marine sandstone deposited in a shallow epicratonic embayment on the Texas platform (Goolsby, 1957; Cornish, 1975; Krause, 1996; Teran, 2007). The thickness of the member reaches up to 168 m in paleotopographic lows (Barnes and Bell, 1977). The Hickory Sandstone is subdivided into upper, middle, and lower subunits (Fig. 2). The lower subunit was interpreted to be braided stream deposits grading upward into tidal flat and intertidal estuarine deposits, the middle subunit consists of fluvial-influenced, shallow subtidal estuarine and

Figure 1. (A) Map showing the main base-ment provinces and study area in North America. Fort Worth Basin is shown in a stippled pattern. Modified after Sims and Petermar (1986), Mueller et al. (2002), and Dickinson and Gehrels (2009). (B) Map showing tectonic provinces and distribution of surface exposures of Precambrian and Cambrian rocks in Texas. Modified after Thomas (2006), Stoesser et al. (2007), and Whitmeyer and Karlstrom (2007). (C) Geological map of the Fort Worth Basin. Modified after Barnes (1992) North America stages follow Rohde (2005).

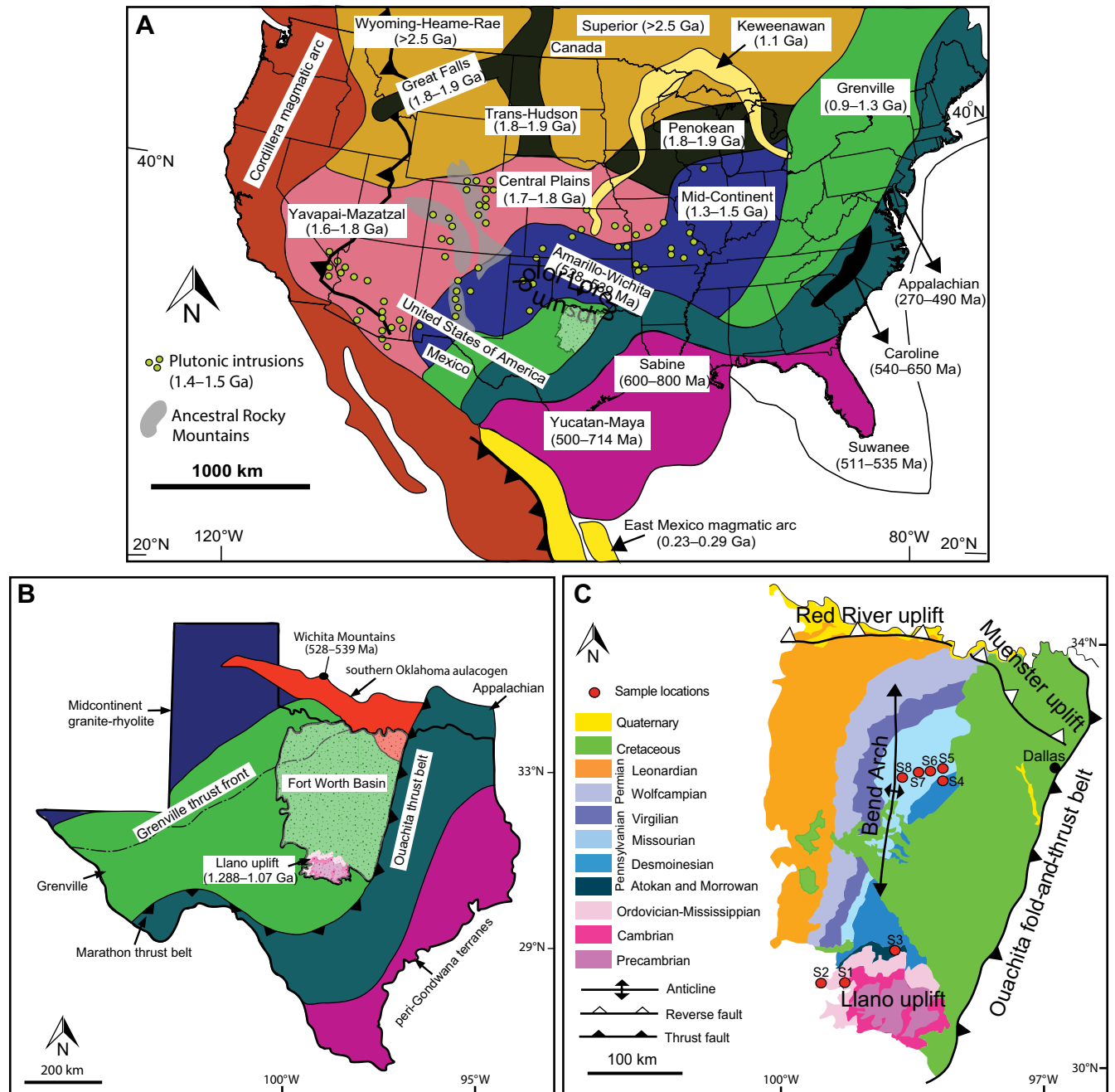


TABLE 1. LOCATION AND SAMPLE DESCRIPTION OF THE SANDSTONE SAMPLES USED FOR DETRITAL ZIRCON GEOCHRONOLOGY IN THIS STUDY, FORT WORTH BASIN, USA

Sample no.	Sample name	Formation (Group)	Epoch	Latitude (°N)	Longitude (°W)	Elevation (m)	Sample description	References
S1	Lower Hickory	Riley (Moore Hollow)	Late Middle Cambrian	30.8371	99.1171	491	Cross-bedded, coarse- to medium-grained, angular to subangular, poorly sorted quartzose sandstone	Bridge et al., 1947; Barnes and Bell, 1977
S2	Upper Hickory	Riley (Moore Hollow)	Late Cambrian	30.7653	99.4051	521	Coarse-grained, moderately to well-sorted, well-rounded quartzose sandstone with iron-oxide ooids and cement	Barnes and Schofield, 1964; Randolph, 1991; Wilson, 2001
S3	Big Saline	Big Saline (Bend)	Middle Pennsylvanian	31.1377	98.5795	397	Medium-grained, subangular quartzose sandstone	Flippin, 1982
S4	Dobbs Valley	Grindstone Creek (Strawn)	Middle Pennsylvanian	32.7050	98.0688	234	Large-scale trough cross-bedded sandstone	Brown et al., 1973
S5	Brazos River	Brazos River (Strawn)	Middle Pennsylvanian	32.7554	98.0562	285	Small-scale trough cross-bedded, ripple cross-laminated, or planar-laminated sandstone	Brown et al., 1973
S6	Lake Pinto	Mineral Wells (Strawn)	Middle Pennsylvanian	32.7828	98.1802	270	Cross-bedded, coarse-grained sandstone	Plummer and Hornberger, 1935
S7	Turkey Creek	Mineral Wells (Strawn)	Late Pennsylvanian	32.7679	98.3125	325	Cross-bedded, coarse-grained sandstone	Plummer and Hornberger, 1935
S8	Colony Creek	Colony Creek Shale (Canyon)	Late Pennsylvanian	32.7529	98.5197	386	Cross-bedded or massive, fine-grained sandstone	Brown et al., 1973

shoreface deposits, and the upper subunit consists of estuarine–shallow marine deposits (Goolsby, 1957; Cornish, 1975; Krause, 1996). Primary depositional cross-beds within the Hickory Sandstone suggest that sediments were mostly transported from north-northwest to south-southeast (Wilson, 1962; Cornish, 1975; Krause, 1996; Cook, 2009). Our samples were collected from the lower and upper subunits.

One Pennsylvanian sandstone sample (S3) was collected from the Big Saline Formation in the southwestern part of the Fort Worth Basin (Fig. 1C; Table 1). Cheney (1940) and Turner (1957) proposed that the Big Saline is a formation within the Middle Pennsylvanian Bend Group, and that the formation overlies the Early Pennsylvanian Marble Falls Formation (Fig. 2). The formation consists of predominantly limestone with thin beds of deltaic sandstone and mudstone in the southwestern part of the basin, and becomes interbedded with sandstone and conglomerate of the Bend conglomerate in the northern part of the basin (Turner, 1957; Thompson, 1982; Flippin, 1982). Lithofacies distribution of the Lower Pennsylvanian Bend Group suggests that the paleorivers flowed from the Ouachita orogen and Muenster uplift toward the northwest and southwest (Hentz et al., 2012).

The other five Pennsylvanian samples (S4–S8) were collected from the northwestern part of the Fort Worth Basin (Fig. 1C; Table 1). Four of the samples were collected from different sandstone members of the upper Middle Pennsylvanian–lower Upper Pennsylvanian Strawn Group, and one sample was from an unnamed sandstone unit in the Upper Pennsylvanian Canyon Group. The Strawn Group consists of interbedded marine limestone and mudstone, and deltaic conglomerate and sandstone (Brown et al., 1973). The lower Strawn Group contains fan deltas and shallow marine deposits, and the upper Strawn Group contains deltaic-fluvial deposits (Brown et al., 1973). These rocks thicken eastward and reach up to 1370 m thick in front of the Ouachita

orogen (Alsaalem et al., 2017). Maps of fractional volume of sand in the Strawn and Canyon Groups suggest that the sandstones were most likely sourced from east-northeast of the basin (Brown et al., 1973).

**■ METHODS**

Zircon crystals were extracted from samples by traditional methods of crushing and grinding, followed by separation with a Frantz magnetic separator and heavy liquids following the method of Dickinson and Gehrels (2008). U-Pb analyses of ~100 single zircon grains per sample were conducted by laser ablation–multi-collector–inductively coupled plasma mass spectrometry (LA-MC-ICPMS) at the Arizona LaserChron Center (Tucson, Arizona, USA; see Supplemental Data<sup>1</sup>). The analyses involved ablation of zircon with a Photon Machine Analyte G2 excimer laser using a spot diameter of 35 μm. For samples with smaller grain sizes, the laser beam size was reduced to 25 μm in diameter. Zircon grains were randomly selected for analysis to avoid bias in size or shape. The ablated material was carried in helium gas into the plasma source of a Nu HR ICPMS, which was equipped with a flight tube of sufficient width that U, Th, and Pb isotopes were measured simultaneously. All measurements were made in static mode, using Faraday detectors with 3 × 10<sup>11</sup> Ω resistors for <sup>238</sup>U, <sup>232</sup>Th, and <sup>208-207-206</sup>Pb, and discrete dynode ion counters for <sup>204</sup>Pb and <sup>202</sup>Hg. Ion yields were ~0.8 mV per ppm. Each analysis consisted of one 15 s integration on peaks with the laser off (for backgrounds), fifteen 1 s integrations with the laser firing, and a 30 s delay to purge the previous sample and prepare for the next analysis. The ablation pit was ~15 μm in depth.

Common Pb correction was accomplished by measuring <sup>206</sup>Pb/<sup>204</sup>Pb and assuming an initial Pb isotopic composition from Stacey and Kramers (1975), and uncertainties of 1.0% for <sup>206</sup>Pb/<sup>204</sup>Pb and 0.3% for <sup>207</sup>Pb/<sup>204</sup>Pb. Subtraction of

<sup>1</sup>Supplemental Data. U-Pb geochronological analyses by laser ablation–multi-collector–inductively coupled plasma mass spectrometry. Please visit <http://doi.org/10.1130/GES01480.S1> or the full-text article on [www.gsapubs.org](http://www.gsapubs.org) to view the Supplemental Data.

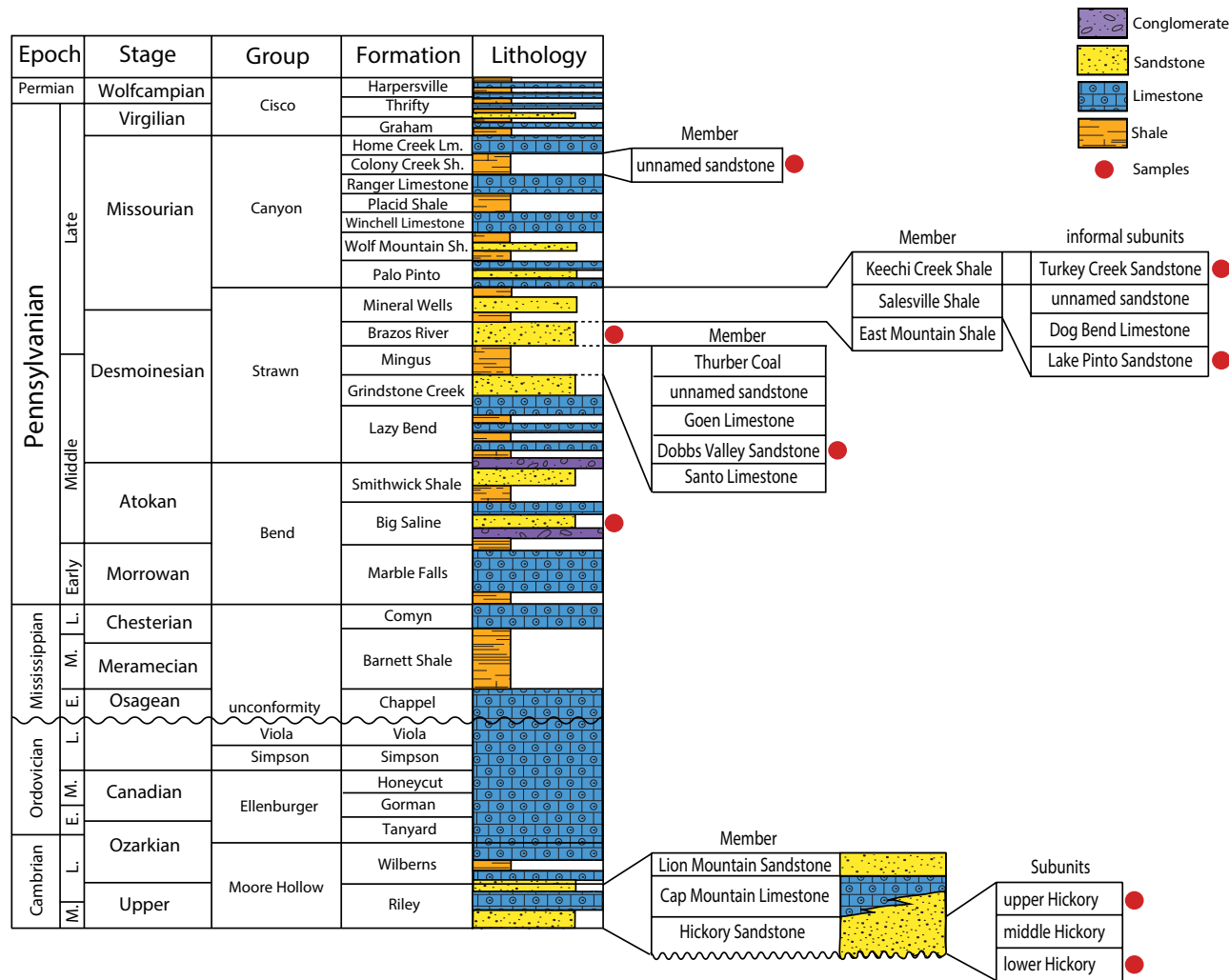


Figure 2. Generalized stratigraphic column of the Fort Worth Basin. Modified after Brown et al. (1973), Flippin (1982), Krause (1996), Pollastro (2003), and Hill et al. (2007). North America stages follow Rohde (2005). E., M., L.—Early, Middle, Late; Lm.—Limestone; Sh.—Shale.

$^{204}\text{Hg}$  was accomplished by using the measured  $^{202}\text{Hg}$  and natural  $^{202}\text{Hg}/^{204}\text{Hg}$  ratio (4.34). Uncertainty in this value of  $^{202}\text{Hg}/^{204}\text{Hg}$  is not significant because of the low intensities of Hg observed. Fractionation of  $^{206}\text{Pb}/^{238}\text{U}$  and  $^{206}\text{Pb}/^{207}\text{Pb}$  during ablation was monitored by analyzing fragments of a large concordant zircon crystal standard from Sri Lanka that has a known age of  $563.5 \pm 3.2$  Ma (Gehrels et al., 2008). The uncertainty arising from this calibration correction, combined with the uncertainty from decay constants, age of the primary standard, and common Pb isotopic composition, contributes 1% systematic error

to the  $^{206}\text{Pb}/^{238}\text{U}$  and  $^{206}\text{Pb}/^{207}\text{Pb}$  ages ( $2\sigma$ ). R33 was used as a secondary standard to ensure that data were reliable. For all samples reported herein, the average  $^{206}\text{Pb}/^{238}\text{U}$  age of the R33 analyses was within 2% of the known age.

The  $^{207}\text{Pb}/^{206}\text{Pb}$  ages are used for grains older than 900 Ma, and  $^{206}\text{Pb}/^{238}\text{U}$  ages are used for grains younger than 900 Ma. Grains with ages older than 400 Ma were filtered by 20% discordance and 5% reverse discordance. Age groups were determined by identifying three or more grains with overlapping  $^{206}\text{Pb}/^{238}\text{U}$  and  $^{207}\text{Pb}/^{206}\text{Pb}$  ages in the aggregate data set. After filtering

the data for discordance, normalized relative age probability diagrams were constructed from least-discordant data (Fig. 3). Quantitative comparisons of detrital zircon ages were conducted using the Kolmogorov-Smirnov (K-S) test (Press et al., 1986). The test compares two age distributions to determine if they are from the same sources by measuring the probability ( $P$ ). A  $P$ -value  $< 0.05$  rejects the null hypothesis that the two samples are statistically indistinguishable. However, the K-S test results might imply different sources when, for example, the two samples are dominated by the same two age populations but in slightly varying proportions (e.g., Dickinson et al., 2010). Therefore, visual comparison of age spectra was used along with the K-S test.

## RESULTS

The detrital zircon U-Pb dates range from Archean (3.635 Ga) to Pennsylvanian (318 Ma) (Supplemental Data [footnote 1]). As would be expected from their different tectonic setting, the two Cambrian samples have different age distributions compared to the Pennsylvanian samples, and the distributions of the six Pennsylvanian samples are similar to each other. Because the Cambrian samples show relatively simple age distributions compared to the Pennsylvanian samples, here we classify zircon populations based on the Pennsylvanian samples. Based on known ages of major magmatic terranes in North America (e.g., Thomas et al., 2004; Dickinson and Gehrels, 2009; Soreghan et al., 2002; Gehrels et al., 2011) and the abundance of zircons from each terrane in our samples, a total of 1002 dates are divided into five populations (Fig. 3): Archean–Paleoproterozoic (3.634–1.825 Ga) zircons of population A, late Paleoproterozoic–early Mesoproterozoic (1.825–1.300 Ga) zircons of population B, middle Mesoproterozoic–early Neoproterozoic (1.300–0.900 Ga) zircons of population C, Neoproterozoic–earliest Paleozoic (800–500 Ma) zircons of population D, and Paleozoic (500–318 Ma) zircons of population E. The six Pennsylvanian samples include 11.5% population A, 15.0% population B, 46.8% population C, 15.3% population D, and 10.3% population E zircons. The remaining 1.1% of the zircons are of 900–800 Ma age. Given that zircons of this age group do not have well-defined magmatic terranes in Laurentia and the content is very low in our samples, they are excluded from our discussion. The two Cambrian samples have 79.9% population B, 17.7% population C, and 2.4% population D zircons. Zircons of population B in the two Cambrian samples are in a tight age range of 1.451–1.325 Ga. Zircons of population D are only present in the upper subunit of the Hickory Sandstone, and have a tight age range of 535–508 Ma.

## POTENTIAL ZIRCON SOURCES

### Archean and Paleoproterozoic (>1.825 Ga)

Potential sources of zircons from population A (3.634–1.825 Ga) mainly include the Archean Wyoming-Heame-Rae and Superior provinces (3.600–2.500 Ga), the Paleoproterozoic Penokean–Trans-Hudson provinces (2.000–1.800 Ga), and the Great Falls tectonic zone (ca. 1.86 Ga) in the interior of Lau-

rentia (Fig. 1A) (e.g., Hoffman, 1988; Frost et al., 2000; Mueller et al., 2002; Whitmeyer and Karlstrom, 2007). Although these sources were distal to the Fort Worth Basin and were not exhumed during the Paleozoic, zircons of this population have been found in a wide range of Proterozoic to lower Paleozoic strata of North America (e.g., Stewart et al., 2001; Gehrels et al., 2011; LaMaskin et al., 2011; May et al., 2013), particularly in the lower Paleozoic strata in southern Laurentia, including the Ordovician–Lower Pennsylvanian strata in the Appalachian foreland basins (Eriksson et al., 2004; Park et al., 2010); Middle Ordovician–Pennsylvanian strata in the Ouachita foreland basins in Oklahoma and Arkansas (Sharrah, 2006; Pickell, 2012); late Mesoproterozoic strata in western Texas (Spencer et al., 2014); and Cambrian–Ordovician strata in southern New Mexico (Gleason et al., 2007; Amato and Mack, 2012). About 2.1% of zircons in the Pennsylvanian samples are of 2.300–2.000 Ga age. This subgroup of zircons was originally formed by the Trans-Amazonian–Eburian orogenic events in the Gondwana margin (Trompette, 2000), and has been found in the Devonian–Pennsylvanian strata in the Appalachian foreland basin, reflecting recycling from various peri-Gondwana terranes (Thomas et al., 2004; Park et al., 2010; Mueller et al., 2014).

### Late Paleoproterozoic–Early Mesoproterozoic (1.825–1.300 Ga)

Zircons of population B (1.825–1.300 Ga) were formed in the Yavapai–Mazatzal orogenic belt (1.825–1.600 Ga) and the granite-rhyolite province (1.585–1.300 Ga) distributed in the southwestern and midcontinent regions of North America (Fig. 1A). The granite-rhyolite province was formed by anorogenic magmatism (e.g., Hoffman, 1988; Van Schmus et al., 1996; Barnes et al., 2002). Bickford et al. (2015) suggested that the magmatism occurred across the continent with a westward-younging trend during 1.5–1.4 Ga, and that younger magmatism (1.390–1.340 Ga) was a major event only in the south-central midcontinent. The late Paleoproterozoic–early Mesoproterozoic basement was exhumed by a Transcontinental Arch during the Cambrian–Mississippian, and the arch extended from Minnesota southwestward to New Mexico (Billo, 1985; Sloss, 1988). The basement was also exhumed by the ARM orogeny during the Pennsylvanian–early Permian in Colorado, Arizona, New Mexico, and Oklahoma (e.g., Kluth, 1986). Zircons of this population represent a significant component in Mesoproterozoic–lower Permian strata in western USA (Stewart et al., 2001; Dickinson and Gehrels, 2003; Gehrels et al., 2011; Amato and Mack, 2012). A small amount of zircons of this population was also found in the Middle Ordovician–Pennsylvanian siliciclastic rocks in Oklahoma and Arkansas (Sharrah, 2006; Pickell, 2012) and the Neoproterozoic–Pennsylvanian strata in the Appalachian foreland basins (Eriksson et al., 2004; Park et al., 2010).

### Mesoproterozoic to Early Neoproterozoic (1.300–0.900 Ga)

Zircons of population C (1.300–0.900 Ga) were formed during the Grenville orogeny when several continent-continent collisions formed the supercontinent Rodinia (e.g., Dalziel, 1997; Hoffman, 1991; Borg and DePaolo,

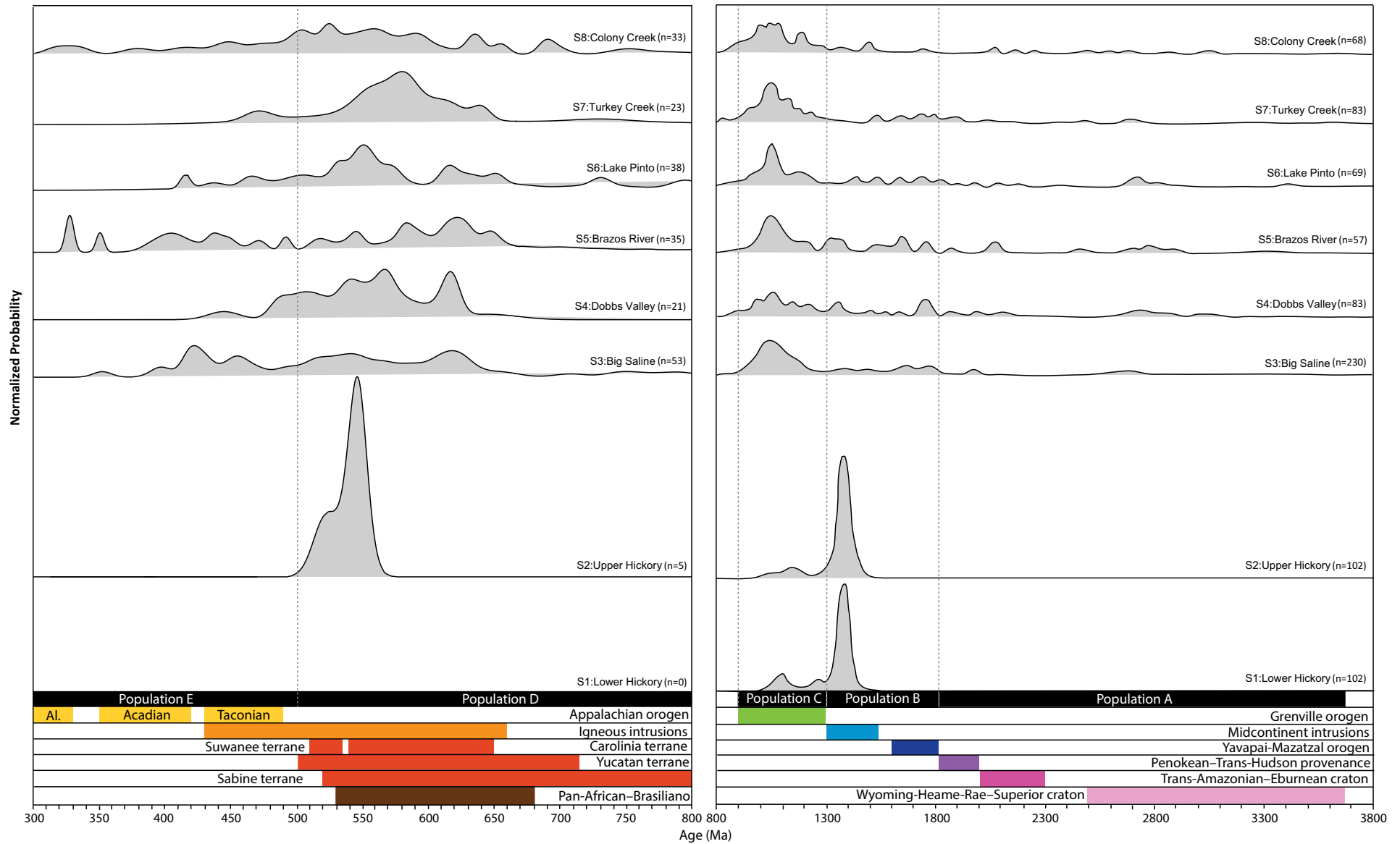


Figure 3. Normalized detrital zircon U-Pb age probability plots for the two Cambrian and six Pennsylvanian sandstones of the Fort Worth Basin. Different provinces and durations of zircon sub-populations A–E are shown at the bottom. Al.—Alleghanian. n denotes the number of concordant detrital zircon ages obtained for each sample.

1994; Torsvik et al., 1996; Unrug, 1995; Whitmeyer and Karlstrom, 2007). The Grenville basement was located in the eastern margin of Laurentia during the Paleozoic, and currently extends from Fennoscandia through Canada to the southeastern USA and forms much of the west Texas and eastern New Mexico basement (Mezger et al., 1993; Dalziel, 1997; Moecher and Samson, 2006; Barnes et al., 2002). Grenville basement was also redistributed to the Precordillera region in South America during the early Paleozoic (Thomas et al., 2002). Zircons of Grenvillian age are very fertile and have been documented as a predominant population in many Neoproterozoic to Paleozoic strata in southeastern Laurentia, particularly in the Appalachian foreland basins (Eriksson et al., 2004; Becker et al., 2005; Moecher and Samson, 2006; Gleason et al., 2007); the Ouachita Mountains in Oklahoma and Arkansas (Gleason et al., 2002); and the Precambrian rocks in west Texas (Shannon et al., 1997; Mosher, 1998; Bickford et al., 2000; Grimes and Copeland, 2004; Spencer et al., 2014).

### Neoproterozoic–Earliest Paleozoic (800–500 Ma)

Zircons of population D (800–500 Ma) were formed in two major tectonomagmatic units with overlapping ages: the peri-Gondwana terranes (800 to ca. 500 Ma), and magmatism on the eastern and southern margins of Laurentia associated with the breakup of Rodinia (760–550 Ma). The major peri-Gondwana terranes (800–500 Ma) include the Avalonian-Carolinian-Uchee terranes (650–540 Ma) in the southern Appalachian Mountains (Rast and Skehan, 1983; Williams and Hatcher, 1982; Murphy et al., 1992; Mueller et al., 1994; Keppie et al., 1996; Thompson et al., 1996; Steltenpohl et al., 2008); the Suwannee terrane (535–511 Ma) in the Florida subsurface (Opdyke et al., 1987; Mueller et al., 1994; Murphy et al., 2004; Martens et al., 2010); the Sabine terrane (800–600 Ma; Thomas, 2013) in eastern Texas and the western Louisiana subsurface (Granata, 1963; Gleason et al., 2007; Nunn, 2012); and the Yucatan-Maya terrane (714–500 Ma) in southeastern Mexico (Rankin et al., 1989; Mueller et al., 1994; Thomas et al., 2004). These terranes were accreted to Laurentia before and during the Appalachian-Ouachita orogenies (ca. 490–260 Ma) (Samson et al., 2001; Murphy et al., 2004; Mueller et al., 2014).

The breakup of Rodinia was associated with two magmatic events in southeastern Laurentia, including a failed rift event at 760–700 Ma and the opening of the Iapetus Ocean at 620–550 Ma (Hatcher, 1989; Su et al., 1994; Aleinikoff et al., 1995; Rankin et al., 1997; Walsh and Aleinikoff, 1999; Cawood and Nemchin, 2001). In the vicinity of the Fort Worth Basin, granitic bodies of ca. 539–528 Ma age were formed in the southern Oklahoma aulacogen (Lambert et al., 1988; Hames et al., 1995; Hogan and Gilbert, 1998; Spencer et al., 2014; Thomas et al., 2016) and exposed in the Amarillo-Wichita uplift and Arbuckle Mountains during the early Paleozoic (Johnson et al., 1988; Gilbert and Denison, 1993). Igneous intrusions related to the opening of the Iapetus Ocean or extension of the southern Oklahoma aulacogen were also found in New Mexico and southern Colorado (Hogan and Gilbert, 1998; Cawood and Nemchin, 2001; McMillan and McLemore, 2004) and were exhumed during the Pennsylvanian–early Permian ARM orogeny (Ye et al., 1996; Dickinson and Lawton, 2003). Small

amounts of zircon of population D (800–500 Ma) have been documented in Paleozoic strata in the Grand Canyon region of Arizona (Gehrels et al., 2011), and in lower Paleozoic strata in southeastern Laurentia, including in the Appalachian foreland basins and the Arkoma Basin (Eriksson et al., 2004; Becker et al., 2005; Moecher and Samson, 2006; Sharrah, 2006; Gleason et al., 2007).

### Paleozoic (500–318 Ma)

Zircons of population E (500–318 Ma) were mainly formed during the Appalachian orogeny (ca. 490–270 Ma). The tectonomagmatic units of the Appalachian orogeny that occurred in eastern Laurentia include the Taconic (490–430 Ma), Acadian (420–350 Ma), and Alleghanian (330–270 Ma) orogenies (Hatcher, 1989; Miller et al., 2000). Zircons of this population have been documented in lower Paleozoic strata in the Appalachian foreland basins (Eriksson et al., 2004; Thomas et al., 2004; Park et al., 2010), Mississippian–lower Permian strata in the Grand Canyon region (Gehrels et al., 2011), and Ordovician–Pennsylvanian strata in Oklahoma and Arkansas (Gleason et al., 2002; Sharrah, 2006). Some grains in this population are of Mississippian age and were derived from volcanic arcs formed during the subduction of southern Laurentia beneath Gondwana. Evidence of volcanic activity of Mississippian age has been documented in sedimentary rocks in Oklahoma and Arkansas (Niem, 1977; Shaulis et al., 2012) and in the Marathon uplift (Imoto and McBride, 1990).

## PROVENANCE INTERPRETATION

The two Cambrian samples have 79.9% population B, 17.7% population C, and 2.4% population D zircons. The predominance of zircons from the granite-rhyolite province of population B suggests that the Cambrian paleorivers mainly drained the granite-rhyolite province. The Grenvillian zircons of population C may be directly derived from the Grenville basement or recycled from older sedimentary rocks exposed in the drainage of the paleorivers. Zircons of 535–508 Ma (population D) were most likely directly sourced from the southern Oklahoma aulacogen (Fig. 3).

The six Pennsylvanian samples include 11.5% population A, 15.0% population B, 46.8% population C, 15.3% population D, and 10.3% of population E zircons. Zircons of the Archean–Paleoproterozoic population and the late Paleoproterozoic–early Mesoproterozoic population (populations A and B) were recycled from Neoproterozoic–Paleozoic sedimentary rocks on the southern margin of Laurentia. Some grains of the late Paleoproterozoic–early Mesoproterozoic population may be directly derived from the ARM region in western Laurentia. The Grenvillian zircons (population C) were directly derived from the Grenville basement, which was exhumed in the Appalachians during the late Paleozoic, or recycled from the Neoproterozoic–Paleozoic sedimentary rocks on the southern margin of Laurentia. The Grenvillian zircons account for nearly half of the grains, suggesting that they were most likely directly derived from the Grenville basement by a major river. Zircons of the Neoproterozoic–earliest Paleozoic population (population D) were derived from the peri-Gond-

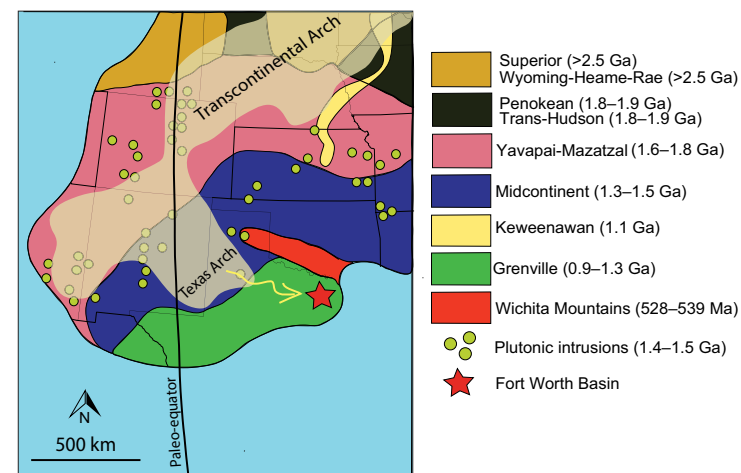
wana terranes (800–500 Ma) and the basin-bounding Amarillo-Wichita uplift and Arbuckle Mountains (ca. 539–528 Ma). The closest peri-Gondwana terrane to the Fort Worth Basin is the Sabine terrane located in the subsurface of east Texas and Louisiana (Fig. 1). Because our Pennsylvanian samples contain only eight grains of 539–528 Ma age, the local ARM terranes are not a major source of zircons of this population. The Paleozoic zircons (population E) were recycled from Paleozoic strata or directly transported from the Appalachians by a transcontinental river. Only a few zircons in this population are of Mississippian age, which were recycled from the Mississippian sedimentary rocks in Oklahoma and Arkansas.

## DISCUSSION

### Sediment Dispersal during the Cambrian

We suggest that the abundant zircons derived from the granite-rhyolite province were transported by paleorivers draining the Texas Arch during the Cambrian (Fig. 4). The Texas Arch was a Cambro-Ordovician structural high on the flanks of the larger transcontinental arc (Fig. 4) near the study area (e.g., Adams, 1954; Wright, 1979; Billo, 1985), and may have remained as a positive topographic feature during the Early Mississippian (Ruppel, 1985). Uplift of the Transcontinental Arch exhumed both the Yavapai-Mazatzal and granite-rhyolite provinces distributed in the southwestern and midcontinent of North America (e.g., Hoffman, 1988; Van Schmus et al., 1993; Dickinson and Gehrels, 2009). Because the granite-rhyolite province was mostly to the south and east of the Yavapai-Mazatzal province, our data suggest that the paleorivers had headwaters only on the south limb of the Texas Arch. Our inference is consistent with previous interpretations that the lower subunit of the Hickory Sandstone was deposited in braided rivers flowing toward the southeast based on stratigraphic architecture, paleocurrent directions, and sandstone compositions (Wilson, 1962; Cornish, 1975; Krause, 1996; McBride et al., 2002). Our interpretation is further supported by the observation that these zircons are in a tight age range (1.451–1.325 Ga), indicating that they were from southwestern Laurentia because the Proterozoic anorogenic magmatism has a westward-younging trend (Bickford et al., 2015). Although the upper subunit of the Hickory Sandstone was deposited in an estuarine–shallow marine environment, the similarity of zircon age distributions of the two Cambrian samples suggests that the grains were mainly transported by fluvial process, and that other depositional processes, such as longshore drift, did not influence the zircon populations.

We suggest that the Grenvillian zircons were directly sourced from Grenville basement exposed on the south limb of the Texas Arch (Fig. 4). The Llano uplift, which bounds the Fort Worth Basin to the south (Fig. 1), is the closest source for the Grenvillian zircons. However, basin subsidence modeling suggests that the Llano uplift was not exhumed until the Ouachita orogeny during the late Paleozoic (Erlach and Coleman, 2005). Based on detrital zircon geochronology, Spencer et al. (2014) proposed that the Llano uplift was the



**Figure 4.** Paleogeographic map showing basement provinces in southern Laurentia and our inferred southeastward sediment dispersal (yellow) from the Texas Arch (light brown) to the Fort Worth Basin (red star) during the Late Cambrian. Basement province map is modified after Thomas (2006), Stoesser et al. (2007), and Whitmeyer and Karlstrom (2007).

main source of Grenvillian zircon to west Texas during the Cambrian. This observation is in contrast to the results of our Cambrian samples. Only 17.7% of zircons are of Grenvillian age, and the ages of these zircons are 1.295–1.040 Ga. The Llano uplift contains Grenville basement (1.360–1.232 Ga) and Proterozoic plutons (1.288–1.070 Ga) (Mosher, 1998). If the Llano uplift was the source for the Grenvillian zircons, we would expect the ages of the Grenvillian zircons in the Fort Worth Basin to be 1.360–1.070 Ga. However, the age range of the Grenvillian zircons in the two Cambrian samples is tight, and the ages are in the younger end of the basement age in the Llano uplift. The Grenvillian zircons were not recycled from older sedimentary strata because our samples do not contain any zircons from the Yavapai-Mazatzal and Archean basement provinces (Fig. 3). Proterozoic sedimentary rocks, such as in west Texas (Spencer et al., 2014), have zircons from the Grenville, Yavapai-Mazatzal, and Archean basements. Recycling of the older sedimentary rocks should not contribute only Grenvillian zircons.

### Sediment Dispersal during the Pennsylvanian

The high abundance (47%) of Grenvillian zircons in the Pennsylvanian sandstones cannot be simply explained by recycling from Proterozoic–lower Paleozoic strata. Although abundant Grenvillian zircons have been found in the Mesoproterozoic and Cambrian rocks in west Texas and lower Paleozoic strata in southern Laurentia (Fig. 5A; Gleason et al., 2002; Gehrels et al., 2011; Amato and Mack, 2012; Spencer et al., 2014), *P*-values of K-S tests conducted on all zircon grains older than 900 Ma in our samples versus these Proterozoic–

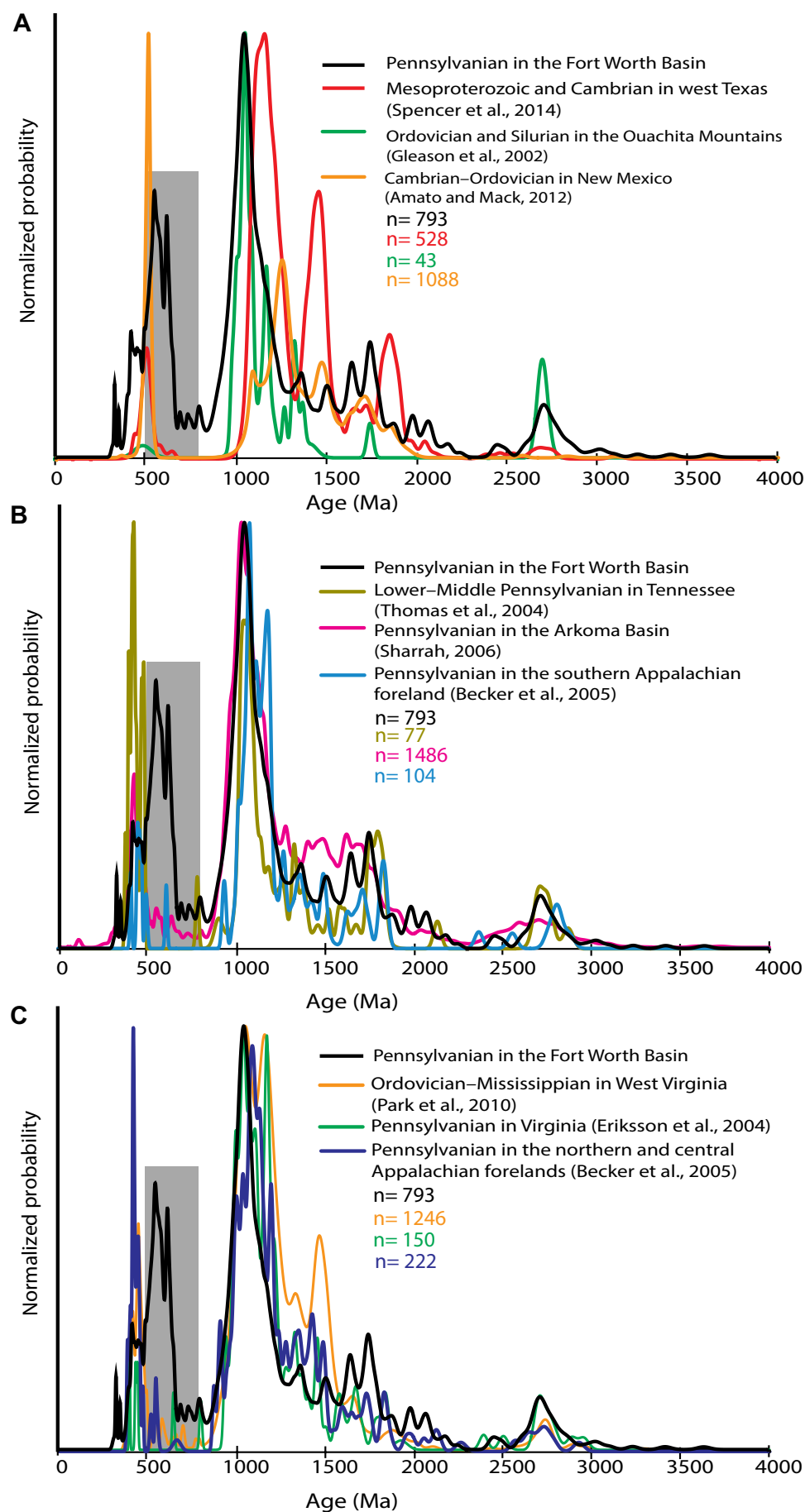


Figure 5. Normalized U-Pb age probability plots comparing our data to those of other studies. (A) Our Pennsylvanian samples in the Fort Worth Basin and other Mesoproterozoic–lower Paleozoic strata in southern Laurentia. (B) Our Pennsylvanian samples in the Fort Worth Basin and other Paleozoic strata in the Delaware Basin. (C) Our Pennsylvanian samples in the Fort Worth Basin and lower Paleozoic strata in the Appalachian forelands. Grey areas highlight the differences in detrital zircon population D between our samples and other Paleozoic strata. n is the total number of zircon data presented in each study.

lower Paleozoic strata are  $\leq 0.05$  (Table 2), indicating that it is very unlikely that the zircon populations are from the same sources or that recycling from the older strata is not a major contributor to sediments in the Fort Worth Basin. We conducted K-S tests on only the grains older than 900 Ma because our samples have high abundance of zircon of 680–520 Ma (Fig. 5), which may have been added to the Pennsylvanian river system from a local peri-Gondwana source, explaining the low *P*-values. K-S tests were performed on all grains older than 900 Ma, assuming that no significant amount of these zircons was dispersed from a local peri-Gondwana source. This assumption can be tested only when future study directly constrains the detrital zircon age distribution of the peri-Gondwana terrane.

Previous studies have suggested that a late Paleozoic river flowed along the Appalachian-Ouachita foreland and brought Grenville basement detritus to the remnant ocean basins (Graham et al., 1975; Archer and Greb, 1995; Gleason et al., 1994, 1995, 2007). Others, however, have suggested that the paleoriver or another paleoriver entered the Ouachita foreland during the early Pennsylvanian by draining through the Illinois Basin (Thomas, 1997; Archer and Greb, 1995; Sharrah, 2006). The paleoriver draining the Illinois Basin (Archer and Greb, 1995) would have had its drainage mostly in the Laurentia continental interior rather than the Appalachians, and cannot explain the high abundance of Grenvillian zircons and low abundance of zircons of Archean–early Mesoproterozoic age. Furthermore, comparison of our detrital zircon data and data from the Pennsylvanian sedimentary rocks in the central Appalachian foreland suggests that the paleoriver did not have a major catchment in the central Appalachians. Zircons from the Paleozoic strata in the central Appalachian foreland have two Grenvillian peaks (Fig. 5C; Eriksson et al.,

2004; Becker et al., 2005; Park et al., 2010), including one at 0.980–1.090 Ga representing the Ottawan orogeny (Rivers, 1997; Heumann et al., 2006) and one at 1.160–1.190 Ga representing the Shawinigan orogeny (Chiarenzelli et al., 2010; McLelland et al., 2010). The Grenvillian zircons in our Pennsylvanian samples are mostly in the age range of 0.980–1.090 Ga (Fig. 5), and the low abundance of zircons of 1.160–1.190 Ga in our samples and the Pennsylvanian strata in the Ouachita foreland (Sharrah, 2006) suggest that the central Appalachians and Ouachita foreland were unlikely to have been connected by a paleodrainage. *P*-values of K-S tests conducted on zircon grains older than 900 Ma in our data versus the data of Ordovician–Mississippian strata in Virginia, West Virginia, and Pennsylvania (Eriksson et al., 2004; Becker et al., 2005; Park et al., 2010) are all  $< 0.05$  (Fig. 5C; Table 2), further indicating the difference in zircon provenances.

We suggest that the abundant Grenvillian zircons were directly derived from the southern Appalachian Mountains via a major river (Fig. 6). The zircon populations in our Pennsylvanian samples are very similar to those in the Pennsylvanian sandstones in the Arkoma Basin (Sharrah, 2006). The *P*-value of a K-S test conducted on the two groups of samples is 0.3 (Fig. 5B; Table 2; Sharrah, 2006), suggesting that sediments in the Fort Worth Basin and Ouachita foreland were from the same source, and that the two basins were most likely connected. Although the Amarillo-Wichita uplift induced by the ARM orogeny could have acted as a topographic barrier between the two regions, isopach maps of the Mississippian and Pennsylvanian strata in the Fort Worth Basin show that the Muenster uplift did not cause significant flexural subsidence, and thus was not a major topographic feature during the Pennsylvanian (Alsalem et al., 2017).

TABLE 2. RESULTS OF KOLMOGOROV-SMIRNOV TESTS FOR DETRITAL ZIRCONS OLDER THAN 900 MA IN STRATA OF DIFFERENT AGES SUMMARIZED IN THIS STUDY

	Fort Worth Basin*	West Texas†	Ouachita Mountains‡	Arkoma Basin#	New Mexico**	Tennessee††	West Virginia‡‡	Virginia§§	Central Appalachian foreland***	Southern Appalachian foreland***
Fort Worth Basin		0.000	<b>0.068</b>	<b>0.310</b>	0.000	<b>0.942</b>	0.000	0.016	0.048	0.011
West Texas	0.000		0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ouachita Mountains	<b>0.068</b>	0.000		0.038	0.000	<b>0.209</b>	<b>0.069</b>	<b>0.602</b>	<b>0.099</b>	<b>0.218</b>
Arkoma Basin	<b>0.310</b>	0.000	0.038		0.000	<b>0.645</b>	0.000	0.007	0.026	0.021
New Mexico	0.000	0.000	0.000	0.000		0.000	0.000	0.000	0.000	0.000
Tennessee	<b>0.942</b>	0.000	<b>0.209</b>	<b>0.645</b>	0.000		0.008	<b>0.210</b>	<b>0.524</b>	<b>0.234</b>
West Virginia	0.000	0.000	<b>0.069</b>	0.000	0.000	0.008		<b>0.444</b>	<b>0.320</b>	<b>0.358</b>
Virginia	0.016	0.000	<b>0.602</b>	0.007	0.000	<b>0.210</b>	<b>0.444</b>		<b>0.236</b>	<b>0.851</b>
Central Appalachian foreland	0.048	0.000	<b>0.099</b>	0.026	0.000	<b>0.524</b>	<b>0.320</b>	<b>0.236</b>		<b>0.752</b>
Southern Appalachian foreland	0.011	0.000	<b>0.218</b>	0.021	0.000	<b>0.234</b>	<b>0.358</b>	<b>0.851</b>	<b>0.752</b>	

References:

- \*This study.
- †Spencer et al., 2014.
- ‡Gleason et al., 2002.
- #Sharrah, 2006.
- \*\*Amato and Mack, 2012.
- ††Thomas et al., 2004.
- ‡‡Park et al., 2010.
- §§Eriksson et al., 2004.
- \*\*\*Becker et al., 2005.

Note: P values  $> 0.05$  indicate that two detrital zircon populations are statistically indistinguishable and are highlighted in yellow.

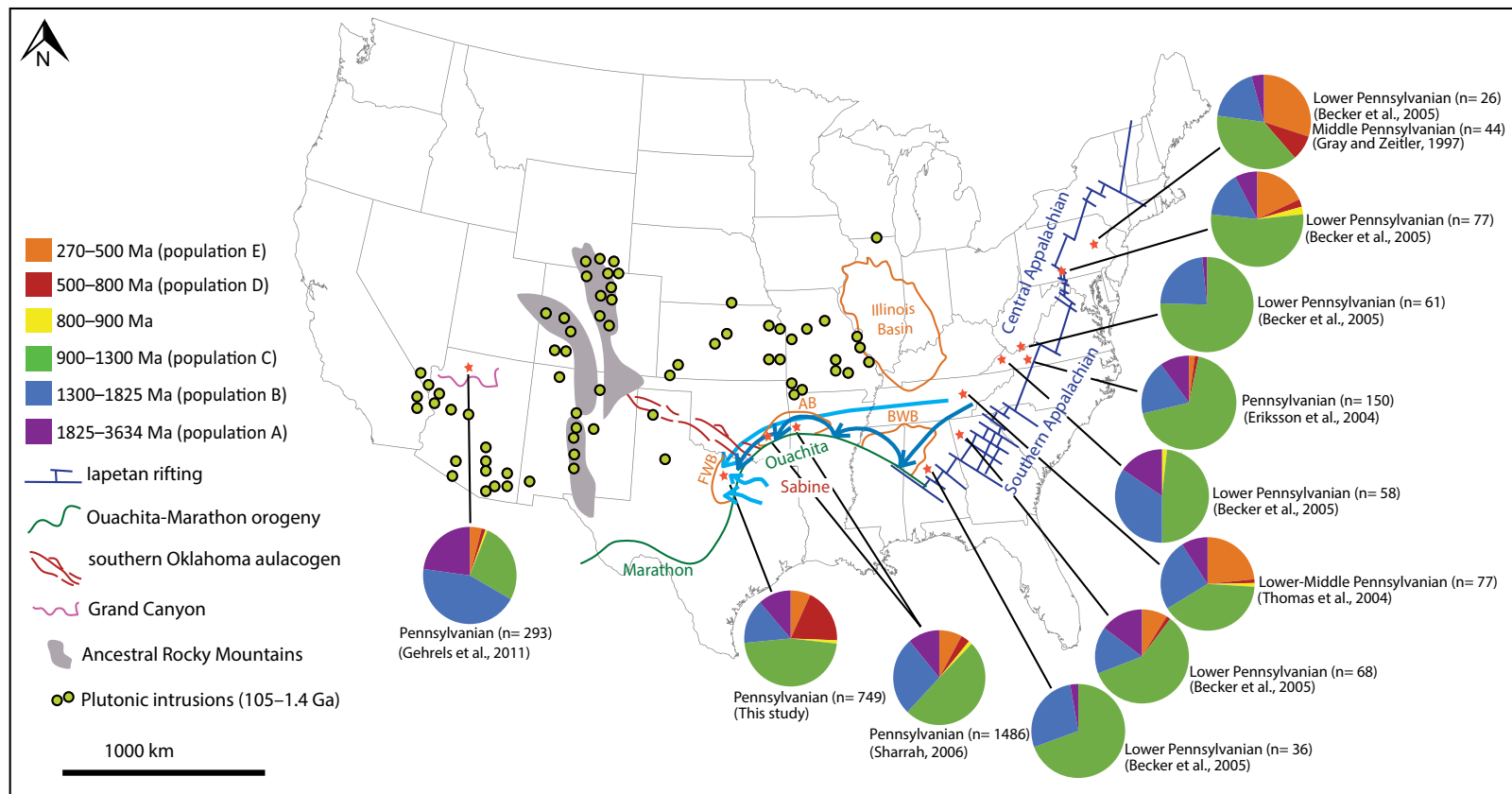


Figure 6. Relative abundance of zircons of populations A–E in southern and eastern Laurentia during the Pennsylvanian and the inferred paleodrainage of the rivers (blue lines) that brought sediments to the Fort Worth Basin. The ages below the pie graphs represent the ages of the strata with detrital zircon studies. n is the total number of zircon data presented in each study. FWB—Fort Worth Basin; AB—Arkoma Basin; BWB—Black Warrior Basin. The long light blue arrow represents our inferred axial river from the southern Appalachians, and the short light blue arrows represent the inferred rivers from the Sabine terrane in the Ouachita orogen. Short arched dark blue arrows represent potential sediment recycling of upper Paleozoic sediments through intermediate steps. See text for details.

The alternative explanation of the abundant Grenvillian zircons is that they were recycled from the Ordovician–Lower Pennsylvanian sandstones that were incorporated in the Alleghanian and Ouachita orogenic fronts when Laurentia and Gondwana experienced oblique collision during the late Paleozoic. Local rivers draining the southern Appalachian orogenic front delivered detritus southward and southwestward and stored the Grenvillian zircons temporarily in the Mississippian–Lower Pennsylvanian sandstones deposited in the remnant Rheic Ocean. During the suturing between Gondwana and Laurentia, these sandstones were incorporated in the orogenic wedge and eroded. Local rivers draining the newly formed orogenic wedge delivered the Grenvillian zircons farther southwestward. This alternative explanation is possible

given that the Grenvillian zircons are abundant in lower Paleozoic rocks and tend to survive recycling. However, we do not prefer this explanation based on three lines of evidence. First, large axial rivers commonly exist in the forelands of long orogenic belts, such as the Ganges River in the Himalayan foreland (Graham et al., 1975). It is very likely such an axial river existed in the southern Appalachian and Ouachita forelands based on the plate tectonic setting. Second, the Mississippian–Lower Pennsylvanian strata deposited in the remnant Rheic Ocean along the Appalachian–Ouachita orogen are dominantly fine-grained siliciclastic sediments and carbonates (e.g., Cline, 1960; Pashin, 1994), thus they are unlikely to represent major sources of coarse sediments in the Fort Worth Basin. The *P*-value of a K-S test conducted on zircon grains

older than 900 Ma in our data versus the data of the Ordovician–Silurian strata in the Ouachita Mountains (Gleason et al., 2002) is only 0.07 (Fig. 5A; Table 2), suggesting that recycling of older strata cannot explain the zircon age distribution in the Fort Worth Basin. Finally, the Mississippian Stanley Group in the Ouachita Mountains has tuffs of 329–321 Ma age (Shaulis et al., 2012), and only one zircon in our Pennsylvanian samples is younger than 330 Ma, suggesting that the flysch deposits in the Ouachita trough were not the major sources of zircons in the Fort Worth Basin.

### Missing Signal of the Amarillo-Wichita Uplift and ARM Orogen

The abundant Grenvillian zircons cannot be derived from the ARM orogen. The Amarillo-Wichita uplift bounding the Fort Worth Basin to the northeast is the nearest ARM uplift to our study site. The uplift has Cambrian igneous rocks of  $535 \pm 10$  Ma and the granite-rhyolite basement of 1.399–1.363 Ga (Thomas et al., 2012, 2016). Erosion of the Amarillo-Wichita uplift should have contributed zircons of these two populations into the Fort Worth Basin, however zircons of  $535 \pm 10$  Ma are present in our Cambrian samples, but not in our Pennsylvanian samples. Zircons of the granite-rhyolite province only contribute a small proportion (~5.6%) to our Pennsylvanian samples. Based on the low abundance, and possible contribution of zircons of the same age from the peri-Gondwana terranes, we suggest that the Amarillo-Wichita uplift did not provide significant amounts of sediment into the Fort Worth Basin during the Middle and Late Pennsylvanian. This interpretation is consistent with the eastward shift of the basin depocenter and dominant control of the Ouachita orogen on basin subsidence during the Middle and Late Pennsylvanian (Alsaalem et al., 2017). The ARM orogeny mainly influenced western Laurentia and reached its maximum during the Middle Pennsylvanian (Kluth and Coney, 1981). The ARM orogeny in western Laurentia exposed the Yavapai-Mazatzal basement, however our samples have only 9.0% zircons of 1.825–1.600 Ga, further suggesting that ARM-involved sources were not a major source of sediments for the Fort Worth Basin during the Pennsylvanian. Recycling of sedimentary cover in the ARM orogen is also unlikely a major source of sediments because the ARM orogen is to the northeast of the Fort Worth Basin and could not have caused westward delta propagation (Brown et al., 1973).

### Signal of the Sabine Uplift

Our Pennsylvanian samples have a zircon age peak at ca. 650–500 Ma and a small proportion of zircons of 650–820 Ma age (Fig. 3). Igneous and metamorphic rocks of these ages have not been documented in the vicinity of the Fort Worth Basin. Zircons of these ages were formed in Gondwana during the Pan-Africa orogeny in Africa and the Brasiliano orogeny in South America, and these orogenies began at ca. 820 Ma and were completed by ca. 500 Ma (Hoffman, 1999; Schaaf et al., 2002). Possible sources of these zircons include

the peri-Gondwana terranes accreted to the Laurentia margin before and during the Ouachita orogeny. Although Neoproterozoic zircons are found in the Avalonia, Carolina, and Suwannee terranes (Opdyke et al., 1987; Murphy et al., 1992, 2004; Mueller et al., 1994), zircons of ca. 650–540 Ma age were absent in the central and southern Appalachian forelands (Fig. 5; Eriksson et al., 2004; Thomas et al., 2004; Park et al., 2010), suggesting that these terranes were not likely sources of the Neoproterozoic zircons in our Pennsylvanian samples. The Yucatan-Maya terranes were not accreted to Laurentia along the Ouachita-Marathon belt until the early Permian (Pindell, 1985; Viele and Thomas, 1989; Dickinson and Lawton, 2001), thus they were also not likely the source of the Neoproterozoic zircons.

The only known peri-Gondwana terrane near the Fort Worth Basin is the Sabine terrane, which exists in the subsurface of east Texas and Louisiana today (Fig. 1). We infer that zircons of ca. 650–500 Ma age in our Pennsylvanian samples were from the Sabine terrane. The interpretation is consistent with the inferred westward delta propagation direction based on fractional volume of sand distribution of the Strawn and Canyon Groups (Brown et al., 1973). Thomas (2013) suggested that the synorogenic sandstones in the Arkoma Basin have a distinctive detrital zircon age group of 800–600 Ma, which may be derived from the Sabine terrane. Our samples also have more zircons of 800–650 Ma compared to the other Pennsylvanian sandstones summarized in Figure 5, supporting the interpretation that local rivers draining the Sabine terrane delivered zircons of 820–500 Ma. Presently, there are no age constraints for basement rocks underlying the upper Paleozoic sedimentary cover in the Sabine terrane. Future dating of the basement rocks can verify our inference about the zircon signal of the Sabine terrane.

## CONCLUSIONS

In this study we utilize detrital zircon U-Pb geochronology to understand sediment dispersal and reconstruct paleogeography on the southern margin of Laurentia before and during the suturing of Laurentia and Gondwana. We suggest that detrital zircons in the Cambrian Hickory Sandstone were transported by local rivers draining the Texas Arch located to the northwest of the Fort Worth Basin and that the Llano uplift was not an active source during the Cambrian. Detrital zircons of the six Pennsylvanian sandstones contain 26% Archean–Mesoproterozoic grains, 47% Grenvillian grains, and 26% Neoproterozoic and early Paleozoic grains. The Archean–Mesoproterozoic grains in these Pennsylvanian sandstones were recycled from the lower Paleozoic strata distributed in southern Laurentia. The high abundance of Grenvillian zircons and the similarity of detrital zircon age distributions between our samples and the Pennsylvanian strata in the Arkoma Basin suggests sediment dispersal by a major river flowing in the Ouachita foreland to the remnant ocean basin. Comparison of detrital zircon age distributions in the Pennsylvanian strata in the Fort Worth and Arkoma Basins with those in the Appalachian forelands suggest that the headwaters of the river was limited to the southern Appala-

chians during the Pennsylvanian. Our Pennsylvanian samples have abundant zircons of 820–500 Ma age, which were transported by local rivers from the Sabine terrane incorporated in the Ouachita orogen. Therefore, both distal sediment dispersal from the southern Appalachians and local sediment dispersal from the Ouachita orogen contributed sedimentation in the Fort Worth Basin during the Pennsylvanian.

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