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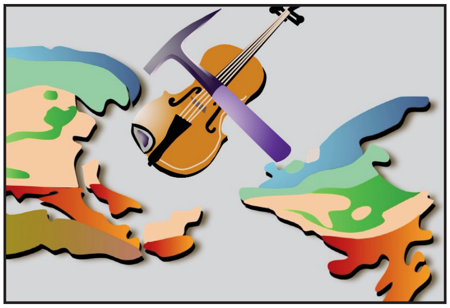

A Mechanism for Tectonic Inheritance at Transform Faults of the Iapetan Margin of Laurentia

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SUMMARY
Transform faults along the Iapetan rifted continental margin of Laurentia offset the continental rift and/or bound domains of oppositely dipping low-angle detachments. Rift-parallel and transform-parallel intracratonic fault systems extend into continental crust inboard from the rifted margin. Ages of synrift igneous rocks, ranging from 765 to 530 Ma, document non-systematic diachroneity of rifting along the Iapetan margin. Synrift sedimentary accumulations show abrupt variations in thickness across transform faults, and some concentrations of synrift igneous rocks are distributed along transform faults and transform-parallel intracratonic fault systems. The greatest thicknesses of Cambrian–Ordovician passive-margin shelf-carbonate deposits are along transform margins and in continental-margin basins along transform faults, as well as along transform-parallel intracratonic fault systems, indicating anomalously great post-rift thermal subsidence along transform faults. Along the Ordovician–Permian Appalachian-Ouachita orogenic belt, a diachronous array of synorogenic clastic wedges fills foreland basins, recording tectonic-load-driven flexural subsidence of the lithosphere. The greatest thicknesses of synorogenic clastic wedges of all ages are consistently in foreland basins along transform margins and inboard from intersections of transform faults with the rifted margin, indicating systematically weaker lithosphere along transform faults. The distinctive and pervasive properties and behaviour of the lithosphere along transform faults in successive tectonic settings suggest fundamental controls on tectonic inheritance at transform faults. Recent models for continental rifting incorporate ductile extension of the mantle lithosphere beneath brittle extension of the crust; the domain of ductile extension of the mantle lithosphere may reach significantly inboard from the rifted margin of the brittle crust, accounting for rift-parallel extensional faults in the crust inboard from the rifted margin. A transform offset of a rift in brittle crust requires a similar offset in ductile extension of the mantle lithosphere, leading to differential ductile flow on opposite sides of the transform and imparting a transform-parallel distributed-shear fabric. Transform-parallel distributed shear in the mantle lithosphere provides a mechanism for brittle transform-parallel fault systems in the continental crust. Studies of seismic anisotropy show fast directions parallel with transform faults, indicating systematic orientation of crystals through transform-parallel distributed shear in the mantle lithosphere.

SOMMAIRE
Les failles transformantes le long de la marge continentale divergente japétienne de la Laurentie décalent le rift continental et/ou les domaines accrétés en des décollements à pendages opposés faibles. Des systèmes de failles intracratoniques parallèles au rift, et parallèles à la transformation, pénètrent vers l'intérieur de la croûte continentale à partir de la marge de rift. Les âges des roches ignées syn-rift, entre 765 Ma et 530 Ma, témoignent d'une activité de rifting diachronique non-systématique le long de la marge japétienne. Des empilements sédimentaires syn-rifts montrent des variations abruptes d'épaisseur d'une faille transformante à l'autre, et des concentrations de roches ignées syn-rifts se répartissent le long des systèmes de failles transformantes et de failles intracratoniques parallèles. Les accumulations les plus épaisses de carbonates de plateforme de marge continentale passive se trouvent le long des marges de cisaillement et dans les bassins de marge continentale le long de failles transformantes, de même qu'au long des systèmes de failles intracratoniques parallèles, évoquant une subsidence anormalement forte le long des failles transformantes. Le long de la bande orogénique ordovicienne-permienne Appalaches-Ouachita,

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une gamme diachronique de prismes clastiques synorogéniques remplit les bassins d'avant-pays, attestant d'une subsidence par flexure lithosphérique d'origine tectonique. Les plus grandes épaisseurs de prismes clastiques synorogéniques à tous les âges sont toujours situées dans les bassins d'avant-pays le long des marges transformantes, et vers l'intérieur, à partir des intersections des failles transformantes avec la marge de rift, indiquant une lithosphère systématiquement plus fragile le long des failles transformantes. Les propriétés particulières et le comportement généralisés de la lithosphère le long des failles transformantes dans les contextes tectoniques successifs sont la marque de contrôles fondamentaux sur l'héritage tectonique des failles transformantes. Les modèles récents de rifting continental comportent une extension ductile de la lithosphère mantellique sous l'extension cassante de la croûte; le domaine d'extension ductile de la lithosphère mantellique peut s'étendre significativement vers l'intérieur de la marge de divergence de la croûte cassante, d'où les failles d'extension parallèle au rift, à l'intérieur de la croûte de la marge de divergence. Un décalage de transformation de rift de la croûte comporte un décalage du même genre de l'extension ductile de la lithosphère mantellique, ce qui implique un différentiel de flux ductile sur les bords opposés de la transformation, d'où cette fabrique d'extension parallèle à la transformation. L'extension parallèle à la transformation de la lithosphère mantellique fournit un mécanisme qui explique les systèmes de failles transformantes parallèles dans la croûte continentale. Les études de l'anisotropie sismique montre les grandes vitesses de propagation parallèles aux failles de transformations, ce qui indique une orientation systématique des cristaux induite par une extension répartie selon les cassures transformantes dans la lithosphère mantellique.

INTRODUCTION

Transform faults along a rifted continental margin offset the continental rift and/or bound upper-plate and lower-plate domains on oppositely dipping low-angle detachments (e.g. Lister et al. 1986; Thomas 1993). Geometry and kinematics require that transform faults are nearly vertical structures, penetrating through the lithosphere to the depth of rift extension. Numerous studies have concluded that transform faults along modern continental margins form abrupt boundaries between domains of contrasting crustal and lithospheric geometries and/or properties (e.g. Isacks et al. 1968; Francheteau and LePichon 1972; Kumar 1978; Sykes 1978; Mascle et al. 1988, 1992; Todd and Keen 1989; Benkhelil et al. 1998). Along-strike variations in Neoproterozoic–Cambrian synrift, Cambrian–Ordovician passive-margin, and Ordovician–Permian Appalachian-Ouachita synorogenic stratigraphy are systematically related to the locations of transform faults of the Iapetan margin of Laurentia (summary in Thomas 2006).

The purpose of this article is to consider Neoproterozoic–Paleozoic distribution patterns along the Iapetan rifted margin of Laurentia in the context of a comprehensive conceptual model for the history of transform faults as a pervasive fabric in the lithosphere. Because the interpretations of transform faults presented here are based on the distributions of synrift sedimentary and igneous rocks, passive-margin shelf-carbonate rocks, and synorogenic clastic wedges, the distribution patterns of each of those are reviewed in some detail. For efficiency in reading, the sections of this article entitled SYNRIFT ROCKS, PASSIVE-MARGIN SHELF-CARBONATE ROCKS, and SYNOROGENIC CLASTIC WEDGES are each organized into three components, beginning with a synthesis entitled Overview and Summary to provide context, including a map to illustrate the distribution patterns, and ending with more detailed descriptions and reference citations to document the synthesis and map.

IAPETAN RIFT-STAGE STRUCTURES

Along the Iapetan continental margin of Laurentia, northwest-striking transform faults offset northeast-striking rift segments (Fig. 1) (Thomas 1977, 2006). The present locations of elements of the continental margin are arrayed within and beneath the late Paleozoic Appalachian-Ouachita orogenic belt, which reflects closing of oceans and assembly of supercontinent Pangaea (e.g. Thomas 2006). Parts of the continental margin have been dismembered and translated by Appalachian-Ouachita thrust faults, requiring palinspastic restoration for resolution of the original trace of the margin; other parts of the margin have remained in the footwall beneath Appalachian and Ouachita allochthons, requiring subsurface resolution of the trace of the margin. The interpreted original trace of the continental margin, as well as the location and identity of transform faults, reflects assembly of a variety of data, which have been detailed in previous publications and will not be repeated here (e.g. Thomas 1977, 1991, 1993, 2006, 2011; Goetz and Dickerson 1985; Haworth et al. 1988; Keller et al. 1989; Cawood and Botsford 1991; Mickus and Keller 1992; Thomas and Astini 1996, 1999; Cherichetti et al. 1998; Cawood et al. 2001; Cousineau and Longuépée 2003; Harry et al. 2003; Harry and Londono 2004; Tull and Holm 2005; Allen et al. 2009, 2010). The reconstruction of the rifted margin shown in Figure 1 integrates restorations of Appalachian shortening to palinspastic locations of pre-Iapetan basement rocks (now in external basement massifs), synrift sedimentary and igneous rocks, and sedimentary rocks of the passive-margin shelf and shelf edge, as well as geophysical resolution of the margin of continental crust.

Inboard from the rifted continental margin within Laurentian crust, an array of late synrift intracratonic fault systems includes elements that are parallel with rift segments of the margin and others that are parallel with transform faults of the margin (Fig. 1) (summaries in Thomas 1993, 2006, 2010). The rift-parallel intracratonic fault systems generally include sediment-filled grabens, indicating rift-sense extension within continental crust. Some transform-parallel intracratonic fault systems are primarily fracture systems filled with synrift igneous rocks; others include sediment-filled, fault-bounded pull-apart basins.

The transmission of extensional strain into continental crust inboard from the rifted continental margin poses a mechanical question. If the rifted margin reflects brittle detachment between two plates during continental rifting and breakup, what is the mechanism for propagation of extensional strain into the continental crust inboard from the brittle rift? Similarly, how is transform motion transmitted into the crust inboard from the rifted margin? Presumably the latter question is merged with the former if extensional strains are partitioned along strike at transform (transfer) faults, and both rift-parallel and transform-parallel intracratonic faults are generated in the same strain field.

SYNRIFT ROCKS

Overview and Summary

Relationships of distribution of synrift rocks to structures of the rifted margin (e.g. Lister et al. 1986, 1991) provide one of the criteria integrated with others for reconstruction of the outline of the rifted margin, including the locations of transform faults. In turn, the reconstruction of the rifted margin provides a framework for an internally consistent, interdependent interpretation of the effects of transform faults on the distribution of synrift rocks. Thick synrift sedimentary accumulations along the Iapetan rifted margin of Laurentia are interpreted to be the fill of listric half-grabens and other fault-bounded basins in lower-plate settings (e.g. Lister et al. 1986, 1991; Thomas 1993). In contrast, synrift sedimentary deposits generally are very thin or lacking along upper-plate margins, where transgressive passive-margin facies directly overlie basement rocks. Transform faults localize abrupt along-rift changes in synrift stratigraphy. Locally thick synrift successions are concentrated at intersections of transform faults with the rift margin, suggesting maximum synrift subsidence. Synrift igneous rocks are distributed along the rifted margin, and some volcanic systems end along the margin at transform faults. Some igneous rocks are distributed along transform faults. Ages of synrift sedimentary accumulations are constrained by interlayered or superposed igneous rocks and by the ages of the transgressive passive-margin cover deposits. The transition in stratigraphy from synrift to passive margin occurred approximately at the beginning of the Cambrian along much of the Iapetan margin of Laurentia, but later along some segments of the margin.

Late synrift sedimentary successions along intracratonic fault systems include the fill of extensional rift-parallel grabens and the fill of pull-apart basins in a transform-parallel graben system. Igneous rocks dominate the synrift accumulations in a transform-parallel fault system and are distributed in dikes and plutons along other transform-parallel fault systems. Both very thick sedimentary successions in pull-apart basins and mantle-derived magmas indicate steep crust-penetrating fractures along transform-parallel intracratonic fault systems. In contrast, the shallower rift-parallel graben systems suggest listric extensional faults. Along the intracratonic grabens, post-graben (post-rift) carbonate rocks of middle Late Cambrian age overlap the graben-filling sedimentary accumulations, documenting that intracratonic graben-boundary faults are younger than faults of the Iapetan rifted continental margin of Laurentia.

Iapetan Rifted Continental Margin

Ages of synrift igneous rocks along the Laurentian margin indicate two phases of rifting (Badger and Sinha 1988; Aleinikoff et al. 1995). The older synrift igneous rocks have ages of ~765 to 650 Ma (Fig. 1); however, the early phase of rifting did not result in continental breakup and the evolution of a passive margin (Tollo et al. 2004). Synrift igneous rocks with ages primarily between ~615 and 550 Ma (Fig. 1) document the later phase of synrift magmatism, which was associated with continental breakup and opening of Iapetus. A late stage of synrift magmatism spanned 539 to 530 Ma along the Southern Oklahoma transform-parallel intracratonic fault system in association with late-stage rifting of the Argentine Precordillera from the Ouachita embayment of Laurentia during the Early Cambrian (Thomas and Astini 1996; Thomas et al. 2012; Hanson et al. 2013b). The ages of synrift igneous rocks show no evident systematic time-space migration, indicating episodic activity along a diachronous rift, rather than progressive along-rift propagation of extension.

Synrift volcanic rocks (Catoctin, 572 ± 5 to 564 ± 9 Ma, Aleinikoff et al. 1995) gradually thicken northward along the Virginia promontory toward the New Jersey transform at the corner of the Pennsylvania embayment (Fig. 1). Basaltic components generally dominate the Catoctin; however, rhyolites are progressively more abundant northward (Rankin 1975, 1976) toward the New Jersey transform at the corner of the Pennsylvania embayment, suggesting that bimodal volcanism was concentrated at the transform fault.

In the Tennessee embayment (Fig. 1), a synrift clastic sedimentary succession (Ocoee Supergroup), lacking volcaniclastic components, is >10 km thick and evidently includes the fill of multiple fault-bounded basins (King 1964, 1970; Hadley 1970; Rankin 1975; Rast and Kohles 1986; Rankin et al. 1989). The thick synrift succession ends northeastward along strike at a transform fault (Thomas 1991). An important synrift volcanic centre (Mount Rogers Formation) includes bimodal volcanic rocks (Rankin 1970) with an age of 758 ± 12 Ma (Aleinikoff et al. 1995). The Mount Rogers volcanic rocks and overlying glaciogenic clastic sedimentary rocks (Rankin 1993) are isolated at the corner of the Tennessee embayment along the Tennessee-Virginia transform (Fig. 1). The Grandfather Mountain Formation is a structurally isolated, thick succession of clastic sedimentary rocks and volcanic flows (Fig. 1), the stratigraphically highest of which has an age of 742 ± 2 Ma (Fetter and Goldberg 1995). Nearby, temporally related plutons range in age from ~765 to 728 Ma (Fig. 1) (Su et al. 1994; Fetter and Goldberg 1995; Ownby et al. 2004). The separate, distinctive, thick synrift accumulations suggest rotated half-grabens along a lower-plate rift margin, which ends northeastward at the Virginia-Tennessee transform fault (Thomas 1993). The lower-plate, thick synrift sediment accumulation in the Tennessee embayment ends abruptly southwestward along strike at the Georgia transform (Fig. 1), southwest of which the rift structure is an upper plate with no preserved synrift sedimentary rocks on the Alabama promontory (Tull and Holm 2005).

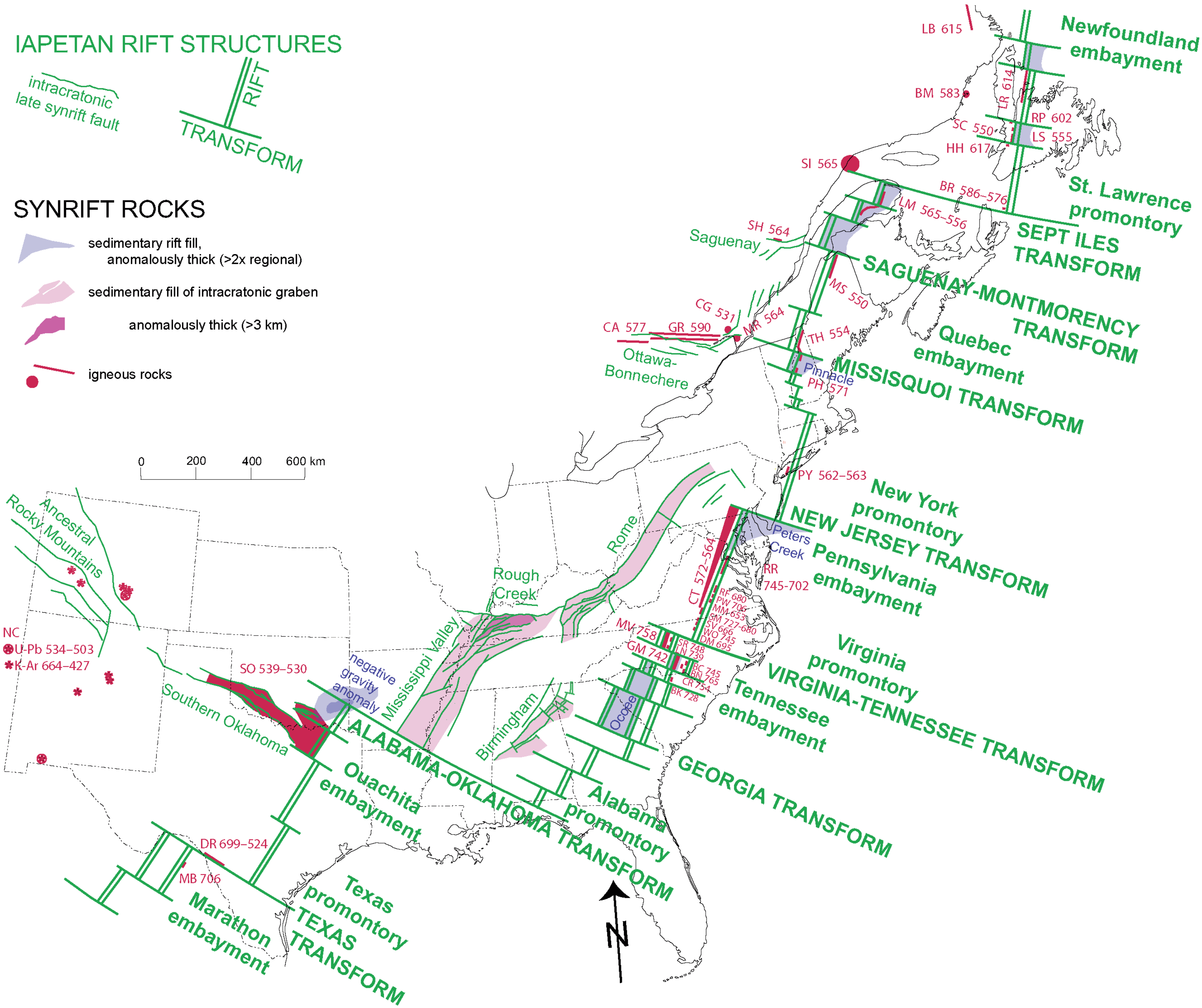
IAPETAN RIFT STRUCTURES
intracratonic late synrift fault
RIFT
TRANSFORM
SYNRIFT ROCKS
sedimentary rift fill, anomalously thick (>2x regional)
sedimentary fill of intracratonic graben
anomalously thick (>3 km)
igneous rocks
0
200
400
600 km
Newfoundland embayment
St. Lawrence promontory
SEPT ILES TRANSFORM
SAGUENAY-MONTMORENCY TRANSFORM
Quebec embayment
MISSISQUOI TRANSFORM
New York promontory
NEW JERSEY TRANSFORM
Pennsylvania embayment
Virginia promontory
VIRGINIA-TENNESSEE TRANSFORM
Tennessee embayment
GEORGIA TRANSFORM
Alabama promontory
ALABAMA-OKLAHOMA TRANSFORM
Ouachita embayment
Texas promontory
TEXAS TRANSFORM
Marathon embayment
LB 615
BM 583
LR 614
RP 602
SC 550
LS 555
HH 617
SI 565
BR 586–576
LM 565–556
SH 564
Saguenay
MS 550
CG 531
MR 564
CA 577
GR 590
Ottawa-Bonnechere
TH 554
Pinnacle
PH 571
PY 562–563
Peters Creek
RR 745-702
CT 572–564
Rome
Rough Creek
MV 758
GM 742
Ocoee
Birmingham
Mississippi Valley
negative gravity anomaly
SO 539–530
Southern Oklahoma
Ancestral Rocky Mountains
NC
U-Pb 534–503
K-Ar 664–427
DR 699–524
MB 706
N

Figure 1. (*facing page*) Outline map of palinspastically restored Iapetan rifted margin of Laurentia, late synrift intracratonic fault systems, and distribution of synrift rocks. Synrift rocks now in Appalachian thrust sheets are shown in interpreted palinspastic locations with respect to the rifted margin. Thick synrift sedimentary accumulations are distributed along lower-plate rift margins and adjacent to transform faults; a large-magnitude negative gravity anomaly at the corner of the Ouachita embayment possibly represents a thick synrift sediment accumulation (Kruger and Keller 1986). Parts of the compilation of ages of synrift igneous rocks were adapted from a compiled map by McCausland et al. (2011) and from an unpublished compilation by Elizabeth McClellan. Reported ages (Ma) of synrift igneous rocks are from U–Pb zircon analyses, except as noted in the following list, which explains the abbreviations: BC—Beech pluton (Su et al. 1994), BK—Bakersville dikes (Ownby et al. 2004), BM—Baie des Moutons syenite ($^{40}Ar–^{39}Ar$, McCausland et al. 2011), BN—Brown Mountain pluton (Fetter and Goldberg 1995), BR—Blair River dikes (Miller and Barr 2004), CA—Callander complex [Ottawa-Bonnechere graben] (Kamo et al. 1995), CG—Chatham-Grenville stock [Ottawa-Bonnechere graben] ($^{40}Ar–^{39}Ar$, McCausland et al. 2007), CR—Crossnore pluton (Su et al. 1994), CT—Catoctin volcanics and dikes (Aleinikoff et al. 1995), DM—Dillons Mill pluton (Fokin 2003), DR—Devils River uplift volcanics (Rb–Sr, Nicholas and Rozendal 1975; Denison et al. 1977; Nicholas and Waddell 1989), GM—Grandfather Mountain rhyolite (Fetter and Goldberg 1995), GR—Grenville dike swarm [Ottawa-Bonnechere graben] (Kamo et al. 1995), HH—Hare Hill granite (van Berkel and Currie 1988), LB—Labrador dikes (Kamo et al. 1989), LM—Lac Matapedia volcanics (Hodych and Cox 2007), LN—Lansing pluton (Su et al. 1994), LR—Long Range dikes ($^{40}Ar–^{39}Ar$, Stukas and Reynolds 1974; Kamo et al. 1989), LS—Lady Slipper pluton (Cawood et al. 1996), MB—Marathon igneous boulder in the Ordovician Ft. Peña Formation [palinspastic location of source not known] (Hanson et al. 2013a), MM—Mobley Mountain pluton (Fokin 2003), MR—Mont Rigaud stock [Ottawa-Bonnechere graben] (Malka et al. 2000), MS—Mount St. Anselme volcanics (Hodych and Cox 2007), MV—Mount Rogers volcanics (Aleinikoff et al. 1995), NC—scattered plutons in New Mexico and Colorado (compilation in McMillan and McLemore 2004), PH—Pinney Hollow metavolcanics (Walsh and Aleinikoff 1999), PW—Polly Wright Cove pluton (Tollo et al. 2004), PY—Pound Ridge granite and Yonkers gneiss (Tollo et al. 2004), RF—Rockfish River pluton (Fokin 2003), RP—Round Pond granite (Williams et al. 1985), RR—Robertson River igneous suite (Lukert and Banks 1984; Tollo and Aleinikoff 1996; Fokin 2003), SC—Skinner Cove volcanics (Cawood et al. 2001), SH—St. Honoré carbonatite complex [Saguenay graben] (K–Ar, Doig and Barton 1968), SI—Sept Isles layered intrusion (Higgins and van Breemen 1998), SM—Suck Mountain pluton (Fokin 2003; Tollo et al. 2004), SO—Southern Oklahoma volcanic-plutonic complex (Wright et al. 1996; Hanson et al. 2009, 2013b; Thomas et al. 2012), SR—Striped Rock pluton (Essex 1992), SV—Stewartsville pluton (Fokin 2003), TH—Tibbit Hill metavolcanics (Kumarapeli et al. 1989), WO—White Oak Creek pluton (Fokin 2003).

The New Jersey transform marks an abrupt along-rift change in synrift stratigraphy. In the Pennsylvania embayment adjacent to the southeastern corner of the New York promontory (Fig. 1), a thick succession of clastic sedimentary rocks (Peters Creek Formation) records synrift marine deposition along the New Jersey transform (Valentino and Gates 1995; Gates and Volkert 2004). North of the New Jersey transform on the southern part of the New York promontory, possible synrift deposits are limited to thin (generally <50 m) accumulations of coarse clastic sediment of the Chestnut Ridge Formation (Gates and Volkert 2004).

Southwest of the Missisquoi transform in the southwesternmost part of the Quebec embayment, alluvial deposits (Pinnacle Formation), as much as 3.5 km thick, are locally interlayered with metabasalt (554 +4/-2 Ma, Tibbit Hill Formation) (Coish et al. 1985; Kumarapeli et al. 1989; Cherichetti et al. 1998). The synrift deposits thin southward away from the transform along the rifted margin and are interlayered locally with volcanic rocks (571 ± 5 Ma, Pinney Hollow Formation) (Stanley and Ratcliffe 1985; Rankin et al. 1989; Walsh and Aleinikoff 1999). The thick synrift deposits of the Pinnacle Formation end abruptly northeastward at the Missisquoi transform; northeast of the transform, the formation is generally <250 m thick (Cherichetti et al. 1998). The thinner Pinnacle shallow-marine clastic deposits overlie bimodal volcanic rocks (~554 Ma, Tibbit Hill Formation) (Marquis and Kumarapeli 1993; Cousineau and Longuépée 2003), indicating synrift deposition later than that southwest of the transform. The abrupt along-strike change in synrift rocks at the Missisquoi transform is interpreted to indicate a change from lower-plate structure on the southwest to upper-plate structure on the northeast (Fig. 1) (Cherichetti et al. 1998; Allen et al. 2010).

Other examples show transform-related variations in synrift stratigraphy. Abrupt along-strike variations in shelf and slope stratigraphy in western Newfoundland indicate that the rift margin on the St. Lawrence promontory was segmented by transform faults (Cawood and Botsford 1991), which separated upper-plate and lower-plate domains along the rifted margin (Allen et al. 2010). Synrift facies coarsen northeastward across the Saguenay-Montmorency transform in the northeastern part of the Quebec embayment (Fig. 1) (Allen et al. 2010). At the corner of the Ouachita embayment, at the intersection of the Alabama-Oklahoma transform with the Ouachita rift (Fig. 1), a very large magnitude negative gravity anomaly suggests a possible deeply buried, thick accumulation of synrift sedimentary rocks (Kruger and Keller 1986).

Intracratonic Fault Systems

Intracratonic grabens (Mississippi Valley, Rough Creek, Rome, Birmingham, Fig. 1) have synrift sedimentary fills of similar compositions and ages; however, the synrift intracratonic graben-fill deposits are younger than the transition from rift to passive margin along the nearest continental rift margins (Thomas 1991). The Mississippi Valley, eastern Rome, and Birmingham grabens are parallel with rift segments of the Iapetan margin (Fig. 1). The Rough Creek and western Rome grabens form an oblique offset in a transform sense from the Mississippi Valley graben to the eastern Rome trough (Fig. 1) (Thomas 1993; Hickman 2011).

Along the Iapetan rifted margin from the Alabama promontory northward to the New York promontory, the transition from rift to passive margin occurred in the earliest Cambrian, as recorded in transgressive passive-margin strata, including a basal sandstone-dominated unit (Chilhowee Group) and an overlying carbonate unit (Shady/Tomstown Dolostone) (summaries in Read, *in* Rankin et al. 1989; Thomas 1991). Extending over much of the southeastern craton above the Shady Dolostone, the upper Lower Cambrian Rome Formation is a fine-grained clastic unit, dominated by redbeds and locally including evaporites (Read, *in* Rankin et al. 1989; Thomas et al. 2001). Detrital zircons indicate sediment supply to the Rome Formation on the Alabama promontory from most of the Precambrian provinces of the Laurentian craton, including the Superior province (Thomas et al. 2004), indicating widespread dispersal of sediment across the passive-margin shelf. Although fault movement may have occurred earlier, Middle Cambrian and lower Upper Cambrian facies and thickness distributions clearly document synsedimentary boundary faults along the intracratonic grabens (Thomas 1991).

Birmingham Graben

Seismically imaged basement faults bound the Birmingham graben beneath the late Paleozoic Appalachian thrust belt on the Alabama promontory (Fig. 1) (Thomas 2007a). The graben fill documented by outcrops and drilling consists of dark-gray mudstones and dark-coloured, fine-grained limestone of the Middle to lower Upper Cambrian Conasauga Formation (Thomas 2001; Thomas and Bayona 2005); the mud-dominated succession grades upward to a carbonate-dominated succession near the top of the formation (Pashin et al. 2012). Palinspastic reconstruction shows the graben fill to be >2 km thick (Thomas 2007a). On both the northwest and southeast shoulders of the Birmingham graben, the lowermost Conasauga Formation of shale and shaly limestone grades upward to massive carbonate rocks, which include intraclastic and oolitic grainstones, indicating shallow-shelf environments adjacent to deeper water in the actively subsiding graben, where slope deposits include debris from the adjacent shelf (Astini et al. 2000). The upward transition from mud-dominated facies to shallow-shelf carbonate facies is much higher stratigraphically in the graben than outside (Thomas et al. 2000). Middle Upper Cambrian carbonate rocks (Knox Group) overstep the graben boundary faults, marking the end of synsedimentary fault movement. The time of initiation of the graben is not documented. Thrust sheets rooted southeast of the Birmingham graben include a thick succession of Rome Formation redbeds and evaporites (Thomas et al. 2001), suggesting another separate graben farther southeast.

Mississippi Valley Graben

Deep drilling and seismic reflection profiles document the boundary faults and sedimentary fill of the Mississippi Valley graben (Fig. 1) beneath the Mississippi Embayment of the Mesozoic–Cenozoic Gulf Coastal Plain (summary in Thomas 1991). The graben contains a Cambrian clastic sedimentary fill that is >1 km thick and is lacking outside the graben. The graben fill consists primarily of fine-grained siliciclastic rocks but includes sporadically distributed fine-grained argillaceous limestone and arkosic to quartzose sandstone in the northwestern part of the graben (Denison 1984; Weaverling 1987; Thomas 1988; Houseknecht 1989). The sandstones suggest alluvial-fan deposition in the graben; the dark-coloured fine-grained rocks in the upper part of the fill indicate deeper water environments. In a shallow fault block on the southeast, the succession includes anhydrite (Mellen 1977). Trilobites indicate an early Late Cambrian age for the upper part of the clastic succession (summary of data in Thomas 1991); however, no biostratigraphic data are available for the older part of the graben fill. Middle Late Cambrian and younger carbonate rocks of the Knox Group overstep the graben boundary faults, indicating the time of cessation of extensional faulting.

Rome Trough

The eastern part of the Rome trough extends northeastward, parallel with the Iapetan rifted margin (Fig. 1) (Shumaker and Wilson 1996). The oldest part of the graben fill consists of sandstone and overlying carbonate rocks interpreted to be, but not biostratigraphically documented as, Early Cambrian in age (Ryder 1992; Gao et al. 2000). A laterally variable graben-filling succession of fine-grained clastic rocks and subordinate limestones generally is ~1200 m thick but locally is as much as 3000 m thick (Gao et al. 2000). Lithofacies boundaries are highly diachronous within the graben (Ryder 1992); fossils from a core of fine-grained clastic rocks indicate a Middle or possibly Late Cambrian age (Donaldson et al. 1975). Carbonate rocks of the Upper Cambrian–Lower Ordovician Knox Group overlap the Rome trough.

Rough Creek Graben

The Rough Creek graben and western part of the Rome trough trend eastward, defining an oblique, dextral, transform-sense offset from the northeast-striking rift-parallel Mississippi Valley graben on the southwest to the northeast-striking rift-parallel eastern Rome trough on the northeast (Fig. 1). The fill of the Rough Creek graben is stratigraphically similar to that in the other grabens, except that it is much thicker at 8 km (Hickman 2011). Faults of the Rough Creek graben generally strike easterly, but curve southwestward at the western ends, describing the geometry of pull-apart basins (Hickman 2011). Eastward along strike and across cross faults, the Rough Creek graben joins the western part of the Rome trough, which extends eastward and curves to the northeast (Fig. 1). Carbonate rocks of the Upper Cambrian–Lower Ordovician Knox Group overlap the fill and boundary faults of the Rough Creek graben. The overall geometry of the Rough Creek graben and the western part of the Rome trough suggests an oblique transform offset of the rift-parallel Mississippi Valley and northeastern Rome grabens (Thomas 1993; Hickman 2011). The transform offset of the intracratonic grabens is parallel and aligned with multiple transform faults of the rifted margin within the Tennessee embayment; however, no continuous basement fault directly links the intracratonic transform-parallel system to the transform faults of the rifted margin (Fig. 1).

Southern Oklahoma Fault System

In the context of the Iapetan rifted margin, the Southern Oklahoma fault system is parallel with transform faults and extends >500 km northwesterly into Laurentian continental crust from the Ouachita embayment in the rifted margin (Fig. 1); the Southern Oklahoma fault system intersects the Ouachita rift margin about 150 km south of the corner of the embayment at the intersection of the rift with the Alabama-Oklahoma transform fault (summaries in Thomas 2010, 2011). The Southern Oklahoma fault system encompasses a suite of bimodal plutonic and volcanic rocks, including gabbro, basalt, granite, and rhyolite, the composition of which indicates mantle sources (Hogan and Gilbert 1998; Hanson et al. 2013b). U–Pb zircon analyses document an age range of 539 ± 5 to 530 ± 1 Ma (Wright et al. 1996; Hanson et al. 2009, 2013b; Thomas et al. 2012). A very large volume (more than 250,000 km^3) of magma was emplaced during that time (Hanson et al. 2013b). High-amplitude, short-wavelength gravity and magnetic anomalies indicate dense mafic rocks with steep boundaries in the shallow continental crust (Keller and Stephenson 2007). Lavas along the Southern Oklahoma fault system are from fissure eruptions, and no volcanic centres are recognized (Hogan and Gilbert 1998). Both the large volume of igneous rocks and overprinting by late Paleozoic large-magnitude basement faults obscure the identity of specific synrift basement faults (Denison, *in* Johnson et al. 1988; Perry 1989). The available ages indicate a short time span for emplacement of the entire igneous complex during the late stages of Iapetan rifting of Laurentia. The geometry and composition of the igneous rocks indicate crust-penetrating, near-vertical fractures as magma conduits, consistent with a leaky transform fault. A transgressive passive-margin succession of basal sandstone and overlying shallow-marine carbonate rocks overlaps the Cambrian igneous rocks, and the age of the base of the transgressive succession is middle Late Cambrian (Denison, *in* Johnson et al. 1988).

Ottawa-Bonnechere Graben

The Ottawa-Bonnechere graben extends into the Laurentian craton parallel with the trend of the Missisquoi transform (Fig. 1). Movement on mappable boundary faults post-dates most Ordovician stratigraphic units, and some synsedimentary fault movement is indicated (Bleeker et al. 2011). Cambrian strata unconformably overlie a low-relief surface on Grenville basement rocks, and no clear record of synrift basement faults is preserved (Bleeker et al. 2011). Dikes of the Grenville dike swarm and Callander complex (~590 to 577 Ma, Kamo et al. 1995) strike approximately parallel with the later graben (Bleeker et al. 2011). In the eastern part of the graben, the Mt. Rigaud stock has a younger age of ~554 Ma (Malka et al. 2000). Lack of evidence of synrift faults suggests that the synrift dikes may have been emplaced along a system of fractures, from which any volcanic edifices were eroded prior to Cambrian deposition (Bleeker et al. 2011).

Saguenay Graben

The Saguenay graben extends into continental crust approximately perpendicular to the strike of the rifted margin in alignment with the Saguenay-Montmorency transform (Fig. 1). Along the graben, two carbonatite complexes are interpreted to be synrift. One of the carbonatites yielded a K–Ar age of 564 Ma (Doig and Barton 1968). The graben contains no recognized synrift sedimentary rocks (e.g. Lavoie and Asselin 1998).

Sept Isles Layered Mafic Intrusion

The Sept Isles transform offsets the rifted margin from the Quebec embayment to the St. Lawrence promontory (Fig. 1). Northwest of the corner of the Quebec embayment along the trend of the Sept Isles transform, the Sept Isles layered mafic intrusion has an age of 564 ± 4 Ma (Higgins and van Breemen 1998). Geochemical and isotopic data indicate a magma source in the upper mantle and a lack of mixing with continental crustal components. The setting suggests magma migration along vertical fracture systems associated with the Sept Isles transform.

Ancestral Rocky Mountains

Within Laurentian crust far from the Iapetan margin, brittle fault-bounded basins of the Ancestral Rocky Mountains generally parallel transform faults along an alignment northwestward from the Alabama-Oklahoma transform of the rifted margin (Fig. 1). Stratigraphic data confirm episodic movement on the basement faults during the late Precambrian through Mississippian, as well as major fault movement during the Pennsylvanian–Permian (e.g. Baars and See 1968; Baars et al. 1988; Thomas and Baars 1995; Thomas 2007b). South of the Ancestral Rocky Mountains, a scattered array of intracratonic plutons with ages of 534 to 503 Ma (U–Pb zircon) and 664 to 427 Ma (K–Ar whole rock and mineral) defines a northeast-trending, rift-parallel, diffuse band (Fig. 1) that extends between the trace of the Southern Oklahoma transform-parallel fault system and the projected trace of the Texas transform, suggesting a failed rift (McMillan and McLemore 2004). Episodic movement on multiple faults of the Ancestral Rocky Mountains includes an important episode during the Cambrian (Baars and See 1968) within the age range of the New Mexico-Colorado pluton swarm. Toward the north, brittle faults of the Ancestral Rocky Mountains diverge from the transform-parallel trend and diminish in magnitude (e.g. Baars et al. 1988).

PASSIVE-MARGIN SHELF-CARBONATE ROCKS

Overview and Summary

A thermal anomaly and crustal uplift associated with high heat flow along an active rift ultimately decay as the rifted continental margin retreats from a spreading mid-ocean ridge; the thickness of the passive-margin carbonate rocks reflects subsidence of the continental shelf in response to decay of the thermal anomaly (e.g. Bond and Kominz 1988; Buck et al. 1988). The passive-margin succession thickens generally toward the continental margin; however, the general pattern of cratonward thinning and transgression of the Cambrian–Ordovician passive-margin deposits along the Iapetan margin of Laurentia (e.g. Sloss 1988) is systematically overprinted by along-margin thickness variations. The thickness of the passive-margin succession is greatest along transform margins with long offsets and in elongate downwarps that plunge toward the rift margin at transform faults (Fig. 2). In addition, transform-parallel intracratonic fault systems are sites of relatively thick passive-margin carbonate rocks, both along a fracture system with igneous rocks and along a fault system with sediment-filled pull-apart basins. The greater thicknesses along transform faults document systematically greater rates of post-rift passive-margin thermal subsidence along the trends of transform faults. These relationships indicate a unique behaviour of post-rift thermal decay of the lithosphere along transform faults at the continental margin and along the intracratonic projections of transform faults. Exponentially decreasing subsidence rates through time characterize thermal decay and cooling of the lithosphere. Subsidence analyses show that post-rift cooling and thermal subsidence of the lithosphere prevailed through the Cambrian and essentially stabilized by the end of the Early Ordovician along the Iapetan margin of Laurentia (Bond et al. 1984; Thomas and Whiting 1995; Thomas and Astini 1999).

Iapetan Rifted Margin

Along the Iapetan margin of Laurentia, a laterally diachronous, transgressive stratigraphic succession above a post-rift unconformity records a transition from active rift to passive margin, and culminates upward in passive-margin carbonate-shelf facies and equivalent off-shelf slope facies (e.g. Thomas 1991; Thomas and Astini 1999; Cawood and Nemchin 2001). The passive-margin succession in eastern Laurentia corresponds to the Sauk Sequence (Sloss 1963), including a basal sandstone, which transgressed cratonward from Early Cambrian to Late Cambrian, and an overlying carbonate-dominated succession, ranging through the Early Ordovician. The craton-wide post-Sauk unconformity provides a useful reference surface for correlation of the top of the passive-margin succession, although the passive margin persisted later along parts of the margin. Continuing synsedimentary extension on late synrift intracratonic fault systems generated local, abrupt thickness variations, as well as a complex interplay of fine-grained clastic and shelf-carbonate facies, until the final overlap of the grabens in the middle Late Cambrian. To avoid those local variations, an analysis of regional patterns of post-rift thermal subsidence here (Fig. 2) relies on the middle Late Cambrian through Early Ordovician deposition of passive-margin carbonate-shelf facies.

Along the southeast side of the Texas promontory adjacent to the Ouachita rift margin, where the basal sandstone is late Middle Cambrian age, the passive-margin carbonate-shelf succession is generally no more than 1000 m thick, reflecting an upper-plate setting and minimal post-rift thermal subsidence (Thomas and Astini 1999). Scattered well data indicate thicknesses generally <1000 m across the Texas promontory (Fig. 2).

Along the Alabama-Oklahoma transform from the Ouachita embayment to the corner of the Alabama promontory, the passive-margin carbonate succession generally is >1500 m thick (Fig. 2). On the west, at the corner of the Ouachita embayment along the Southern Oklahoma fault system, sandstone at the base of the transgressive passive-margin succession is of middle Late Cambrian age (Denison, *in* Johnson et al. 1988). On the east on the Alabama promontory, the base of the passive-margin succession is of Early Cambrian age; however, the succession includes the fine-grained clastic fill of the Birmingham graben (Osborne et al. 2000). The middle Upper Cambrian–Lower Ordovician (Knox Group) passive-margin carbonate-shelf succession overlaps the late synrift faults and graben-fill deposits of the Birmingham basement graben (Thomas 2007a). The clastic facies of the Lower to lower Upper Cambrian (Rome and Conasauga Formations) around the Birmingham graben either pinch out or grade to carbonate facies toward the southwest, and a thick massive shelf-carbonate facies extends along the transform margin (Fig. 2).

Northeast of the Alabama-Oklahoma transform fault on the Alabama promontory (the Alabama arch of Read, *in* Rankin et al. 1989), the passive-margin carbonate-shelf succession thins northeastward to less than 1000 m (Fig. 2). On the Alabama promontory, effects of the Birmingham graben locally modified the preserved thickness of the passive-margin carbonate succession (Knox Group), the upper part of which is unconformably absent in the palinspastically restored cover of the Birmingham graben (Thomas 2007a). Because the original (prior to early Middle Ordovician erosion) thickness distribution with respect to the location of the graben is not known, these effects are omitted from the isopach map in Figure 2.

A locally thick passive-margin carbonate-shelf succession extends

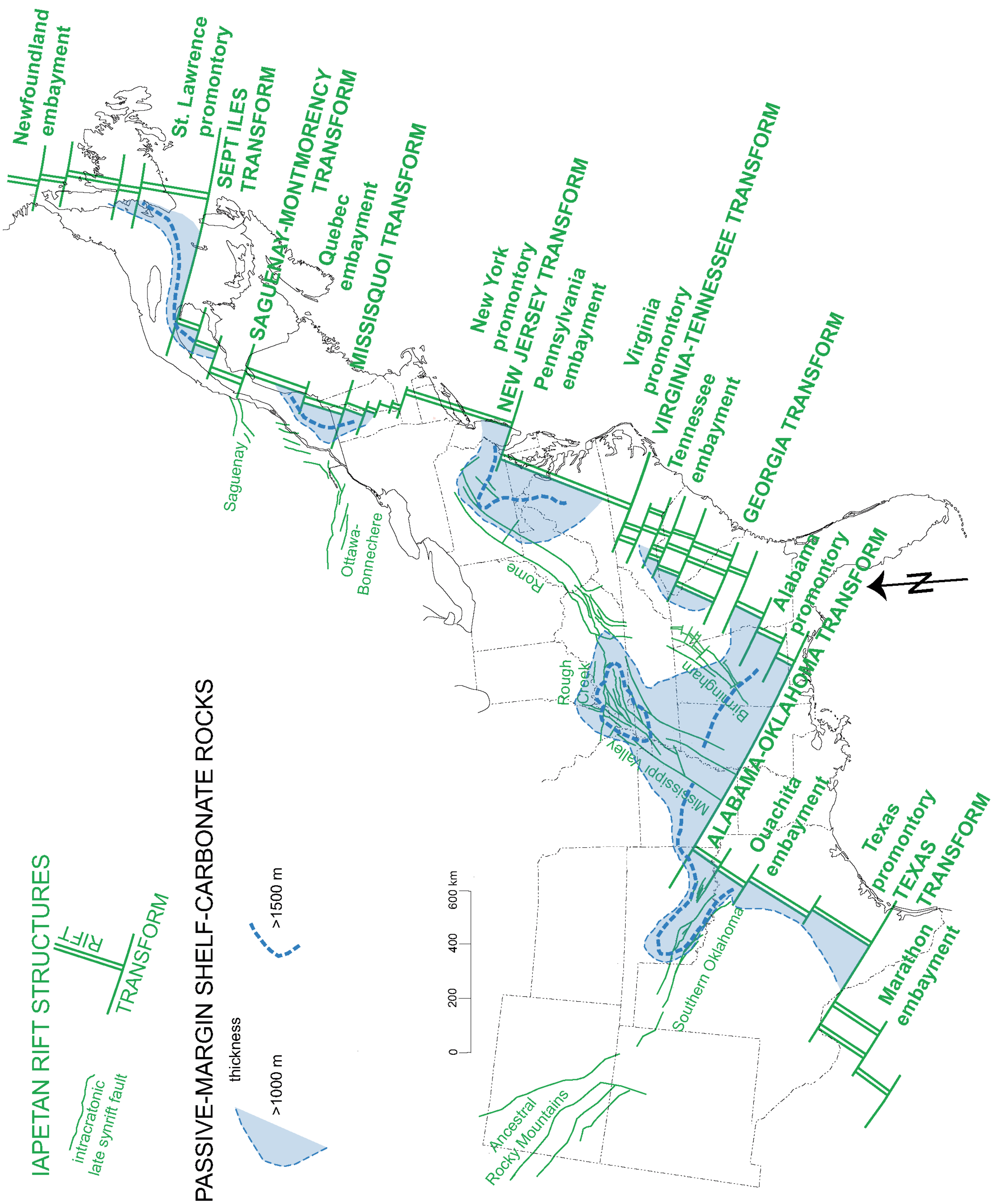

Figure 2. Isopach map of the Upper Cambrian–Lower Ordovician passive-margin shelf-carbonate facies along the Iapetan margin of Laurentia (compiled from Read, *in* Rankin et al. 1989; Sanford 1993; Thomas and Astini 1999; Hickman 2011), showing the association of greatest post-rift thermal subsidence with transform faults of the rifted margin and with transform-parallel intracratonic fault systems.

between the Georgia and Virginia-Tennessee transform faults around the Tennessee embayment (Fig. 2) (Conasauga basin of Read, *in* Rankin et al. 1989). Farther northeast along the rifted margin, the passive-margin carbonate-shelf succession thins to <1000 m on the Virginia promontory (Fig. 2) (Virginia arch of Read, *in* Rankin et al. 1989).

A prominent carbonate depocentre (Pennsylvania depocentre of Read, *in* Rankin et al. 1989) is centred on the Pennsylvania embayment along the New Jersey transform (Fig. 2). The passive-margin carbonate-shelf facies is >1500 m thick in a northwestwardly elongate, southeast-plunging depression, indicating a downwarp extending >200 km inboard from the rifted margin along the trend of the New Jersey transform. Farther north on the New York promontory, the passive-margin succession is uncommonly thin (generally <500 m) (Fig. 2).

Inboard from the Missisquoi transform in the southeastern corner of the Quebec embayment, a thick (>1500 m) passive-margin carbonate-shelf succession fills a southeastwardly plunging depression (Fig. 2) (Sanford 1993). The passive-margin succession thins along the central part of the Quebec embayment between the Missisquoi and Sept Isles transforms. An area of thick passive-margin carbonate-shelf deposits extends along the Sept Isles transform from the northeastern corner of the Quebec embayment to the corner of the St. Lawrence promontory (Fig. 2) (Sanford 1993).

Intracratonic Fault Systems

Along the intracratonic Southern Oklahoma fault system, coincident with the area of synrift igneous rocks, the passive-margin carbonate-shelf succession is exceptionally thick (>1500 m) (Fig. 2) (Denison, *in* Johnson et al. 1988; Thomas and Astini 1999). The base of the transgressive passive-margin succession is a sandstone of middle Late Cambrian age, indicating significant delay in post-rift transgression. The magnitude of post-rift thermal subsidence indicates anomalously great synrift thermal uplift associated with the great volume of synrift igneous rocks, followed by post-rift cooling and subsidence along the transform-parallel Southern Oklahoma fault system (Thomas and Astini 1999).

A broad area of thick (>1000 m) passive-margin carbonate-shelf deposits extends into the craton across the rift-parallel intracratonic Mississippi Valley graben and overlaps the graben-boundary faults and graben-fill sedimentary succession (Thomas 1991). A local area of greater thickness (>1500 m) of passive-margin carbonate-shelf deposits covers the western part of the transform-parallel Rough Creek graben at the offset from the Mississippi Valley graben to the Rome trough (Fig. 2) (Hickman 2011). The carbonate-shelf facies overlaps the graben-boundary faults and the exceptionally thick graben-fill succession of clastic sedimentary rocks. The thickness of passive-margin carbonate-shelf deposits defines a broad downwarp that extends beyond the fault boundaries of the Rough Creek graben and that shows no abrupt thickness variations attributable to synsedimentary fault movement, indicating post-rift passive-margin thermal subsidence along the graben and graben flanks after cessation of late synrift faulting.

SYNOROGENIC CLASTIC WEDGES

Overview and Summary

Multiple episodes of arc collision, terrane accretion, and continental collision constitute the broad spectrum of orogeny that ultimately resulted in the Appalachian-Ouachita orogenic belt and assembly of supercontinent Pangaea (e.g. Hatcher et al. 1989a; Williams 1995). Cratonward verging thrust belts emplaced cratonward tapering wedges as tectonic loads on the edge of Laurentian crust, resulting in flexural subsidence of foreland basins along the margin of the continent. An array of foreland basins, inboard from the continental margin, records flexural subsidence of the lithosphere as a result of tectonic loading through a variety of accretionary events, which span imbrication of Laurentian shelf strata, overthrusting of Laurentian slope deposits onto the continental shelf, accretion of arc terranes and exotic microcontinental terranes, and continental collision.

Synorogenic lithospheric subsidence of a foreland basin reflects emplacement of a tectonic load (the orogen), and the thickness and facies of synorogenic clastic wedges record the magnitude of foreland subsidence and sediment supply (e.g. Jordan 1981; Quinlan and Beaumont 1984). If a foreland basin is ultimately filled, the total thickness of a synorogenic clastic wedge is a measure of the magnitude of foreland subsidence; however, if a basin remains underfilled, receiving deep-water sediment, the total thickness is less than total subsidence. To make a comparison of foreland subsidence between all of the basins, a calculation of subsidence rate takes into account the total thickness of synorogenic sediment in the context of the time span of filling. Figure 3 shows a compilation of subsidence rates for the succession of foreland basins associated with the temporally complex Appalachian-Ouachita orogenic events along the Iapetan margin of Laurentia.

A non-systematic distribution of diachronous events constitutes the whole of the Appalachian and Ouachita orogenies. Classically, events along the orogenic belt have been grouped into the Taconic, Acadian, and Alleghanian/Ouachita orogenies (e.g. King 1959; Hatcher et al. 1989a; Williams 1995). Subsequently, multiple, more temporally and geographically restricted orogenic “events” or “phases” (e.g. Blountian, Salinic, Neoacadian) have been differentiated; however, for purposes of this discussion, broader temporal subdivisions are appropriate to recognize the range of along-strike diachroneity and variability of subsidence along the entire orogenic foreland. Thus, episodes of load-driven lithospheric subsidence and synorogenic filling of foreland basins are grouped simply as Ordovician–Silurian, Devonian–Mississippian, and Mississippian–Permian. Within each of those age groupings, however, the ages of synorogenic filling of foreland basins vary along the continental margin.

Synorogenic foreland basins along the Appalachian-Ouachita orogen share the characteristic of being broadly semicircular in shape. The greatest thicknesses and greatest subsidence rates are located near the centres of the semicircular clastic wedges, with the exception of some basins that are elongate along transform margins (Fig. 3). The shapes of many of the basins

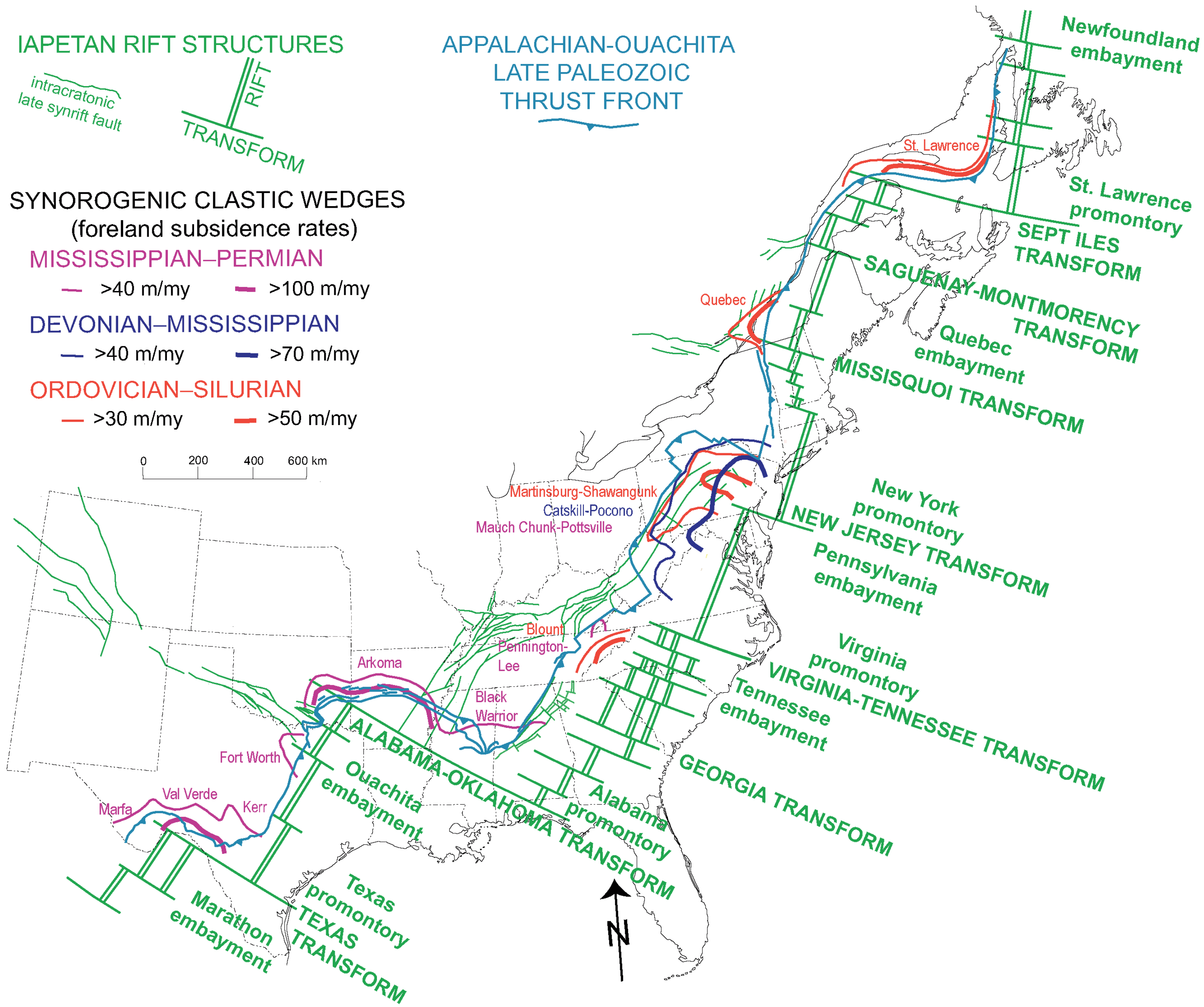

Figure 3. Map of distribution of synorogenic clastic wedges associated with Ordovician–Silurian, Devonian–Mississippian, and Mississippian–Permian orogenic events (expressed as subsidence rate, m/my—metres per million years) (compiled from Thomas 1977; Shanmugam and Walker 1980; Sloss 1988; Sanford 1993; Hiscott 1995). Map shows relation of sites of maximum synorogenic tectonic-load-driven flexural subsidence of foreland basins to the locations of transform faults of the Iapetan rifted margin of Laurentia and to the curved trace of the Appalachian-Ouachita thrust front. Thinner, distal fringes of the synorogenic clastic wedges, indicating lower subsidence rates, extend beyond the mapped outlines of higher subsidence rates, both toward the craton and along-strike of the continental margin.

generally define downwarps that plunge toward the continental margin, where the axes of the basins are aligned with transform faults of the continental margin. The gradual thinning toward the periphery of each basin, both cratonward across orogenic strike and along orogenic strike, indicates a broad downwarp. Some synsedimentary normal faults are generally down toward the basins (e.g. Thomas 2010); however, the margins of the basins generally are not defined by faults. The distribution of facies indicates semiradial progradation of sediment from depocentres near the centres of the semicircular clastic wedges. The thicknesses of the clastic wedges influenced the ultimate shape of the thrust belt, because foreland thrusting propagated farther cratonward in the thickest sedimentary accumulations (Thomas 1977; Marshak 2004).

Along the Appalachian orogenic foreland, the Ordovician–Silurian clastic wedges have maximum thicknesses in the foreland of the Tennessee embayment and along the New Jersey, Missisquoi, and Sept Isles transforms. A Devonian–Mississippian clastic wedge has maximum thickness in the Pennsylvania embayment along the New Jersey transform. Mississippian–Permian clastic wedges have maximum thicknesses in multiple foreland basins along the Alabama-Oklahoma and Texas transforms, and lesser thicknesses in the Tennessee and Pennsylvania embayments (Fig. 3).

The distribution of sites of maximum thickness (depocentres) of clastic wedges indicates a genetic association with transform faults (Thomas 1977). The substantial along-strike variations in magnitude of foreland subsidence require substantial along-strike variations in the cause of subsidence. Generally, foreland subsidence is attributed to some combination of magnitude of the tectonic load and elastic strength of the lithosphere (e.g. Watts and Ryan 1976; Jordan 1981; Karner and Watts 1983; Quinlan and Beaumont 1984). The along-strike variations in foreland subsidence and thickness of synorogenic sediment could be a result of along-strike variations in magnitude of the tectonic load; however, possible along-strike variations in magnitude of the tectonic load (the orogenic hanging wall) have no necessary relationship to characteristics of the loaded margin (the footwall). Instead, the systematic association of centres of greatest magnitude of foreland subsidence with the locations of transform faults at the continental margin suggests that subsidence is controlled by lateral variations in strength of lithosphere systematically associated with transform faults, a suggestion that is enhanced by the multiple occupation of some sites by successive foreland basins. Systematically weaker lithosphere along transforms, in turn, suggests pervasive properties of the lithosphere that are imparted along transform faults during continental rifting.

Ordovician–Silurian Foreland Basins and Synorogenic Clastic Wedges

Synorogenic clastic wedges, ranging in age from early Middle Ordovician to early Late Silurian, record tectonic loading, flexural subsidence of the lithosphere, and sedimentary filling of foreland basins during the initial orogenic events along the Iapetan margin of Laurentia (e.g. King 1959; Thomas 1977; Drake et al. 1989; Williams 1995). Both age and thickness of synorogenic clastic sediment vary significantly along the continental foreland (e.g. Bradley 1989). Ordovician–Silurian foreland basins are distributed along the Iapetan margin of Laurentia from the Newfoundland embayment to the Alabama promontory (Fig. 3). West of the Alabama promontory, passive-margin deposition persisted along the continental margin around the Ouachita and Marathon embayments, indicating no orogenic activity or tectonic loading there during the Ordovician and Silurian (e.g. Viele and Thomas 1989).

Along the southwest side of the St. Lawrence promontory, a lower Middle Ordovician (middle Llanvirn to Llandeilo) synorogenic succession of dark-coloured turbidites thickens southwestward to >1900 m (Sanford 1993; Hiscott 1995) toward the Sept Isles transform (Fig. 3). Farther northwest along the Sept Isles transform at the northeastern corner of the Quebec embayment, an Upper Ordovician (middle Caradoc to lower Ashgill) turbidite succession has a maximum thickness of 4000 m (Hiscott 1995). Maximum subsidence rates are along the Sept Isles transform in the corner of the Quebec embayment and near the corner of the St. Lawrence promontory; however, the initial deposits on the St. Lawrence promontory are older than those in the Quebec embayment. Tectonic loading and subsidence of the shelf, along with progradation of synorogenic sediment, began on the St. Lawrence promontory in the early Middle Ordovician and in the Quebec embayment in the early Late Ordovician (Bradley 1989; Hiscott 1995). Synorogenic turbidites overlie passive-margin slope deposits in the Appalachian allochthons.

In the southwestern corner of the Quebec embayment along the Missisquoi transform, an Upper Ordovician (middle Caradoc to middle Ashgill) dark-coloured turbidite succession thickens generally southeastward toward the continental margin, and has a locally preserved upward gradation to redbeds (Hiscott 1995). The maximum thickness of >1500 m (Sanford 1993) and greatest subsidence rates are along the continental margin along the projected trace of the Missisquoi transform (Fig. 3). Thicker and somewhat older synorogenic turbidites overlie passive-margin slope facies in the Appalachian allochthon; however, tectonic subsidence of the shelf and progradation of synorogenic sediment began along the Quebec embayment in the early Late Ordovician (Hiscott 1995).

The Martinsburg-Shawangunk clastic wedge is thickest (~4000 m) near the corner of the Pennsylvania embayment and exhibits a semicircular distribution of thickness centred on the corner of the embayment (Fig. 3) (Thomas 1977). The maximum subsidence rate at the centre of the semicircular clastic wedge outlines a northwestwardly elongate foreland basin projecting into the craton along the trend of the New Jersey transform in the corner of the Pennsylvania embayment (Fig. 3). An upward gradation from shallow-marine carbonate rocks to dark-coloured shaly limestone indicates initial foreland subsidence in the late Middle Ordovician (Shanmugam and Lash 1982; Drake et al. 1989) [the reported age corresponds to early Late Ordovician of the 2014 International

Chronostratigraphic Chart]. A thick succession of turbidites grades upward and cratonward into Upper Ordovician shallow-marine redbeds (Queenston delta), which spread laterally along the foreland from the centre of maximum thickness (Meckel 1970). In the proximal part of the clastic wedge, an angular unconformity separates the older components from Lower Silurian sandstone, and the clastic wedge is capped by Upper Silurian shallow-marine redbeds and evaporites (summary in Ettensohn 2008).

The Blount clastic wedge is thickest in the centre of the Tennessee embayment near the Virginia-Tennessee transform, where the maximum thickness is >2500 m (Shanmugam and Walker 1978, 1980). An elongate, narrow area of maximum subsidence indicates a relatively high-amplitude short-wavelength foreland basin along the Tennessee embayment (Fig. 3). An upward transition from shallow-marine limestone to black shale records initial foreland subsidence and progradation of synorogenic clastic sediment, beginning in the Middle Ordovician; and the black shale grades upward into a turbidite succession (Shanmugam and Lash 1982). Foreland subsidence began on the Alabama promontory and migrated along the rifted margin into the Tennessee embayment (Bayona and Thomas 2006). In the distal (cratonward) part of the foreland basin, lower Upper Ordovician redbeds and shallow-marine sandstones indicate filling of the basin. In the more proximal part of the basin, the top of the preserved Blount clastic wedge is capped by an unconformity, which is covered by the distal part of the younger Martinsburg-Shawangunk clastic wedge prograding southwestward from the Pennsylvania embayment (Thomas 1977).

Devonian–Mississippian Foreland Basin and Synorogenic Clastic Wedge

The record of a Middle Devonian–Middle Mississippian orogeny in the Appalachian-Ouachita foreland is largely restricted to the area of the Pennsylvania embayment, where the Catskill-Pocono clastic wedge is centred (e.g. King 1959; Thomas 1977; Osberg et al. 1989). The maximum thickness of the clastic wedge is >3500 m (Thomas 1977), and the area of maximum subsidence is centred on the corner of the Pennsylvania embayment at the New Jersey transform (Fig. 3). The initial deposits of the clastic wedge are black shales, which overlie Lower Devonian shallow-marine carbonate rocks (summary in Ettensohn 2008). A cyclic succession of Middle Devonian black shales and muddy turbidites grades upward into Upper Devonian redbeds of the classic Catskill delta (e.g. Meckel 1970; Dennison 1985). Lower Mississippian coal-bearing sandstones and Middle Mississippian redbeds and evaporites prograded cratonward and southwestward along the continental margin over the Catskill redbeds (Thomas and Schenk 1988). Successive depocentres shifted progressively southwestward along the continental margin from the Pennsylvania embayment onto the Virginia promontory in response to dextral transpression of the tectonic load (Ettensohn 1985, 2008; Sevon 1985; Ferrill and Thomas 1988; Thomas and Schenk 1988).

Mississippian–Permian Foreland Basins and Synorogenic Clastic Wedges

Synorogenic clastic sediment spread cratonward from separate foreland dispersal centres in the Pennsylvania and Tennessee embayments in the Late Mississippian to Pennsylvanian and Early Permian (Thomas, *in* Hatcher et al. 1989b). The total thicknesses and subsidence rates of these clastic wedges are of lesser magnitude than those associated with the earlier orogenic events in the same locations. In contrast to the relatively thin clastic wedges in the Pennsylvania and Tennessee embayments, late Paleozoic synorogenic clastic wedges around the Ouachita and Marathon embayments are relatively thick (Fig. 3).

In the Pennsylvania embayment, the Mauch Chunk–Pottsville clastic wedge has a generally semiradial dispersal system from a centre in the corner of the embayment (Thomas 1977). The maximum preserved thickness of the clastic wedge is ~1000 m in the Pennsylvania embayment, and maximum subsidence rates are low (Fig. 3). An Upper Mississippian redbed facies at the base of the clastic wedge progrades over a Mississippian carbonate facies, and the overlying Pennsylvanian succession includes massive quartzose sandstones overlain by a cyclic coal-bearing sandstone-shale succession (Meckel 1970). The relatively wide lateral extent and smaller thickness are consistent with low subsidence rates, a low-amplitude long-wavelength foreland basin, and extensive progradation of fluvial clastic sediment across a filled foreland basin.

In the Tennessee embayment, the Pennington-Lee clastic wedge has a maximum preserved thickness of ~2000 m (Thomas 1977), and subsidence rates are low over most of the foreland (Fig. 3). An Upper Mississippian redbed facies progrades over a Mississippian limestone and extends both southwestward and northeastward along strike (Thomas, *in* Hatcher et al. 1989b). In contrast, the Pennsylvanian sandstone and overlying coal-bearing succession of the Pennington-Lee clastic wedge extend farther northeast and north, and overlap the southwestward prograding components of the Mauch Chunk–Pottsville clastic wedge. The youngest components of the clastic wedge are of Early Permian age (Arkle 1974; Donaldson and Shumaker 1981).

No foreland clastic deposits of Mississippian–Permian age are preserved north of the Pennsylvania embayment. Instead, Late Devonian through Permian deposits of various ages and depositional settings fill multiple, separate pull-apart basins along a system of dextral strike-slip faults along the Appalachian internides from the Newfoundland embayment to the New York promontory (Thomas and Schenk 1988), indicating a transpressional continental collision rather than tectonic loading by overthrusting on the Laurentian margin.

In contrast to limited accumulations of synorogenic sediment along the Appalachian foreland, very thick accumulations of synorogenic sediment characterize both the Ouachita and Marathon embayments. Both embayments include thick muddy to sandy turbidites deposited in off-shelf settings over preorogenic passive-margin slope deposits, followed by thick sandy turbidites deposited in foreland basins formed by subsidence of the continental shelf (e.g. Viele and Thomas 1989).

The earliest synorogenic sediment in the Ouachita embayment consists of Middle Mississippian muddy turbidites deposited over passive-margin slope deposits; off-shelf deposition continued through the Early Pennsylvanian (Arbenz 1989; Viele and Thomas 1989). Foreland subsidence of the continental shelf began in the Middle Pennsylvanian, and was accompanied by deposition of a very thick (>8000 m) synorogenic clastic wedge of sandy turbidites in the Arkoma foreland basin (Arbenz 1989) inboard from the corner of the Ouachita embayment along the Alabama-Oklahoma transform fault. Exceptionally high subsidence rates characterize the Arkoma foreland basin (Fig. 3).

Southeast of the Arkoma basin along the Alabama-Oklahoma transform, foreland subsidence and clastic-wedge progradation in the Black Warrior foreland basin began in the Late Mississippian and continued in the Early Pennsylvanian (Thomas 1988). Maximum subsidence in the Black Warrior basin was less than that in the Arkoma basin (Fig. 3), and the total preserved thickness is ~3000 m (Thomas 1988). The clastic wedge in the Black Warrior basin consists primarily of shallow-marine to deltaic siliciclastic facies. The lower (Mississippian) part of the succession interfingers with shallow-marine carbonate facies up dip to the northeast (Thomas 1988). Computed subsidence rates show a progressive increase through time, characteristic of flexural subsidence of the lithosphere in response to a growing tectonic load (Whiting and Thomas 1994).

Southwest of the corner of the Ouachita embayment along the Ouachita rift margin, subsidence in the Fort Worth basin was less than that in the Arkoma basin, but was greatest adjacent to the intersection of the Southern Oklahoma transform-parallel intracratonic fault system with the rift margin (Fig. 3). After initial deposition of off-shelf muddy turbidites from Middle Mississippian through Early Pennsylvanian, foreland subsidence and sediment progradation onto the shelf led to deposition of ~2000 m of sandy turbidites in the Fort Worth basin (Crosby and Mapel 1975; Kier et al. 1979; Mapel et al. 1979).

In the Marathon embayment (Fig. 1), off-shelf synorogenic deposition began in the Middle Mississippian and continued to early Early Permian over the preorogenic passive-margin slope facies (King 1937; Ross 1986; McBride 1989). Tectonic loading and subsidence of the continental shelf began diachronously along the continental margin around the Marathon embayment, leading to subsidence of the Kerr and Val Verde foreland basins along the northwest-striking Texas transform and the Marfa foreland basin along the northeast-striking Marathon rift margin of the embayment (Fig. 3).

On the southeast along the Texas transform, subsidence of the Kerr basin on the continental shelf began in Middle Pennsylvanian and continued through Late Pennsylvanian (Wright 2011). The maximum thickness of the synorogenic clastic wedge in the foreland basin is >2100 m (Crosby and Mapel 1975), and subsidence rates are highest in the centre of the basin northeast of the Texas transform margin (Fig. 3).

Near the corner of the Marathon embayment, subsidence of the Val Verde basin began in the Late Pennsylvanian, but a thin unit of black shale indicates a low rate of sediment input to a starved basin (Yang and Dorobek 1995; Hamlin 2009). An overlying thick succession of latest Pennsylvanian to Early Permian turbidites indicates continued foreland subsidence and increased sediment supply. The maximum thickness of the clastic wedge is >3000 m (Hamlin 2009); subsidence rates are relatively high along an elongate, high-amplitude short-wavelength foreland basin along and parallel with the Texas transform margin (Fig. 3).

Southwest of the corner of the Marathon embayment, a thick succession of Upper Pennsylvanian to Lower Permian turbidites indicates foreland subsidence of the Marfa basin (Fig. 3) (Ross 1986; Luff and Pearson, *in* Frenzel et al. 1988). The maximum thickness of the synorogenic clastic wedge is ~3000 m (Ross 1986); high subsidence rates characterize the basin near the corner of the Marathon embayment (Fig. 3).

DISCUSSION AND CONCLUSIONS

Observations of synrift structure of both the rifted margin and intracratonic fault systems, distribution of synrift rocks, post-rift passive-margin thermal subsidence, and synorogenic tectonic-load-driven foreland flexural subsidence all focus on unique lithospheric properties and behaviour along transform faults. Rift-parallel and transform-parallel intracratonic faults (Fig. 1) indicate extensional strain and partitions at transforms within continental crust inboard from the brittle detachment at the rifted continental margin. Locally thick synrift sediment and abrupt thinning of synrift sediment across interdependently interpreted locations of transform faults indicate abrupt along-strike variations in magnitude of synrift subsidence across transform faults. Similarly, synrift igneous rocks along transform faults and along transform-parallel intracratonic faults indicate that some transform faults are conduits for magmas from deep lithospheric sources. Locally thick synrift sediment along transform-parallel intracratonic faults (Fig. 1) indicates anomalous magnitude of synrift subsidence in transtensional pull-apart basins. In the thermal evolution of passive margins, systematically greater subsidence along transform margins, along elongate basins projecting into the craton from transform faults at the continental margin, and along transform-parallel intracratonic faults (Fig. 2) indicates systematically greater magnitudes of post-rift thermal subsidence at transform faults than elsewhere along the continental margin. Finally, the systematic distribution of greater magnitudes of synorogenic tectonic-load-driven flexural subsidence of foreland basins (Fig. 3) along transform faults at the continental margin indicates differentially weaker continental lithosphere along transform faults. Together these observations suggest a need to consider the effects of transform faults on the properties of the continental lithosphere in successive, different tectonic regimes.

A comprehensive conceptual model proposed here provides a solution to the question of the role of transform faults in defining some specific properties of the lithosphere (Fig. 4). Previously proposed models for a

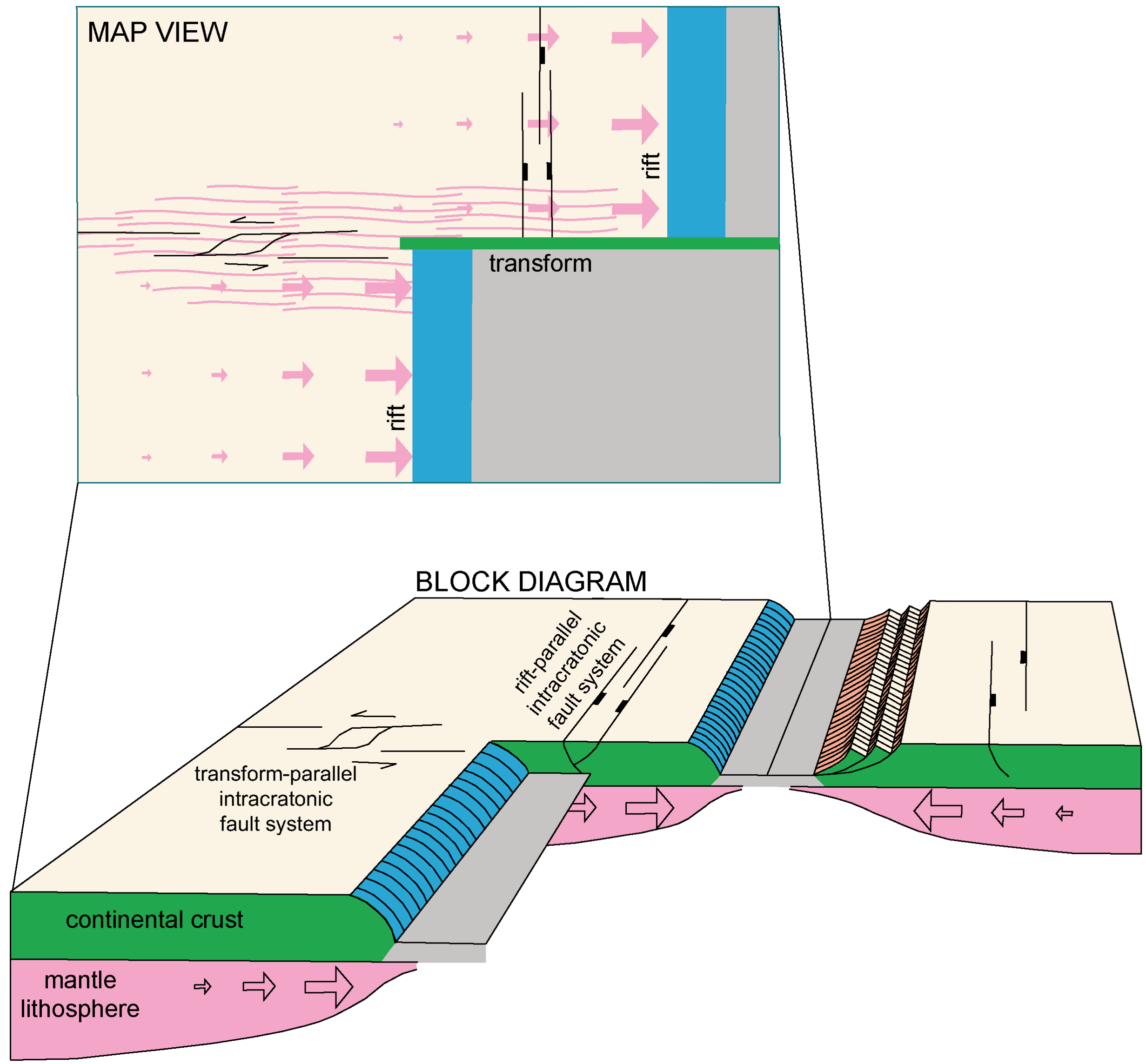

Figure 4. Schematic block diagram and map of conceptual model for continental rifting, adapted from Lister et al. (1986, 1991) for brittle extension and rifting of the crust, and from Lavier and Manatschal (2006) for ductile extension and attenuation of the mantle lithosphere. In addition to a transform fault in the brittle crust, the model incorporates a transform-parallel ductile distributed-shear fabric in the mantle lithosphere. In the block diagram, open arrows show the direction (by arrow heads) and relative magnitude (by size of arrows) of ductile stretching (extension) in the mantle lithosphere; note that, although the lithospheric plates move relatively away from the spreading mid-ocean ridge, stretching (brittle extension in the crust, ductile extension in the mantle lithosphere) is toward the evolving continental margin. The map illustrates crustal structures (rift margin, transform margin, extensional rift-parallel intracratonic fault systems, and transform-parallel intracratonic fault systems), shows direction and relative magnitude (pink arrows) of ductile stretching (extension) in the mantle lithosphere below the crust, and schematically illustrates differential ductile extension of the mantle lithosphere across a transform, which is expressed as a zone of distributed shear (pink wavy lines).

general mechanism of continental rifting include ductile extension of mantle lithosphere beneath brittle extension of the crust, leading to attenuation of the mantle lithosphere and exhumation of the asthenosphere at rifted continental margins (e.g. Lavier and Manatschal 2006). The initial brittle extension of the crust involves listric faults that sole into a low-angle detachment (e.g. Lister et al. 1986), leading to crustal thinning and breakup (separation) at an evolving mid-ocean ridge. Brittle breakup of the continental crust and initiation of a mid-ocean ridge detach the continental crust from the rift-related realm of extensional strain (Fig. 4), thereby leaving no mechanism for later extensional strain within continental crust inboard from the rifted margin to produce the late-stage rift-parallel intracratonic grabens. The domain of ductile extension of the mantle lithosphere, however, must reach some distance beneath the continental crust inboard from the brittle rifted margin of the crust (Fig. 4). Sub-crustal lithospheric ductile extension provides a mechanism for upward propagation of extensional strain into the brittle crust, and offers an explanation for rift-sense brittle extension on rift-parallel intracratonic faults. Extensional movement on the rift-parallel intracratonic faults (primarily Early to early Late Cambrian) is generally the latest stage of rift-sense extension, commonly distinctly younger than the faults of the adjacent rifted continental margin (primarily late Neoproterozoic), consistent with continuing ductile extension of the mantle lithosphere independent of the time of brittle breakup of the crust.

The magnitude of ductile extension of the mantle lithosphere is unconfined laterally and, presumably, is essentially constant along strike at straight segments of the rifted margin. A transform offset of the rift, however, must also form a boundary between domains of ductile extension of the mantle lithosphere. At any point along a transform fault, the mantle lithosphere below the brittle crust on one side of the transform is at a different distance from the rift margin on that side of the transform than is the juxtaposed lithosphere on the other side directly across the transform. Simplistically, the mantle lithosphere on the side of the transform closest to the corresponding rift margin will have a greater rate/magnitude of ductile extension than that on the other (more distant from the rift margin) side of the transform. Thus, a transform in the mantle lithosphere becomes a zone of subhorizontal ductile distributed shear along a diffuse boundary between domains of differential ductile flow on opposite sides of the transform (Fig. 4) (e.g. Baldock and Stern 2005). The transform zone of distributed shear in the mantle lithosphere offers the potential for upward propagation of transform motion into the brittle crust, which may drive transform-parallel fractures and intracratonic faults. Such a mechanism accounts for lack of lateral continuity of transform-parallel intracratonic fractures along strike, because strain can be accommodated in discontinuous transtensional pull-apart basins (e.g. the Rough Creek graben, Fig. 1). The necessary geometry of transform-parallel sub-vertical deep fractures provides a conduit for heat flow and magmas from the mantle lithosphere along "leaky" transform faults (e.g. the Southern Oklahoma fault system, Fig. 1).

Association of transform faults in the crust with distributed-shear zones in the mantle lithosphere offers an explanation for anomalously high heat flow at transform faults. The common association of mantle-derived magmas along intracratonic transform-parallel fault systems, along with anomalously great post-rift thermal subsidence, suggests crust-penetrating steep fractures linked to a ductile-shear fabric in the mantle lithosphere as a conduit for higher heat flow along transforms, particularly leaky transforms. Higher rift-stage heat flow is followed by post-rift thermal decay and greater subsidence in a passive-margin setting along transform faults at the continental margin, inboard from the margin along the projection of transform faults, and along transform-parallel intracratonic faults (Fig. 2). Along the Iapetan margin of Laurentia, post-rift thermal subsidence had essentially stabilized by the end of the Early Ordovician.

Ductile distributed shear in the mantle lithosphere must impart a tectonic fabric along transform faults. If the transform-parallel fabric persists through time, it offers an explanation for systematically weaker lithosphere along transform faults, and a mechanism to explain the concentration of greatest synorogenic tectonic-load-driven foreland flexural subsidence at transform faults (Fig. 3). Tectonic loading and foreland subsidence began diachronously along the Iapetan margin of Laurentia in the Middle Ordovician and continued diachronously and episodically along strike of the Appalachian-Ouachita orogen into the Early Permian.

The proposed conceptual model implies the persistence of a lithospheric fabric parallel with transform faults. Recent observations of seismic anisotropy suggest a way to test the proposed model. Observations along the Alpine fault in New Zealand show a band of seismic anisotropy with fast direction parallel with the Alpine transform fault (Baldock and Stern 2005). The anisotropy is interpreted to result from the alignment of olivine crystals into a pattern parallel with the direction of mantle flow in a zone of distributed shear. Similar observations have been made along other modern transform faults (e.g. Savage and Silver 1993; Herquel et al. 1999; Savage 1999; Rümpker et al. 2003; Roy et al. 2007), and observations of older crust and mantle lithosphere show "fossil" anisotropy (Babuška and Plomerová 2006). A recent compilation of data for seismic anisotropy in North America (Liu 2009) shows that most fast directions in shear-wave splitting are parallel with the modern direction of absolute plate motion, indicating that the source of the anisotropy is in the asthenosphere. Some sites of seismic anisotropy, however, have fast directions that are not parallel with absolute plate motion, suggesting that the source of the anisotropy may be in the mantle lithosphere. A filter of all directions at angles of more than 25° from absolute plate motion shows that most have orientations approximately parallel with transform faults of the Iapetan margin and inboard projections of those transforms (Fig. 5). Most of those sites are concentrated along the inboard projections of transform faults and are not randomly distributed. These examples

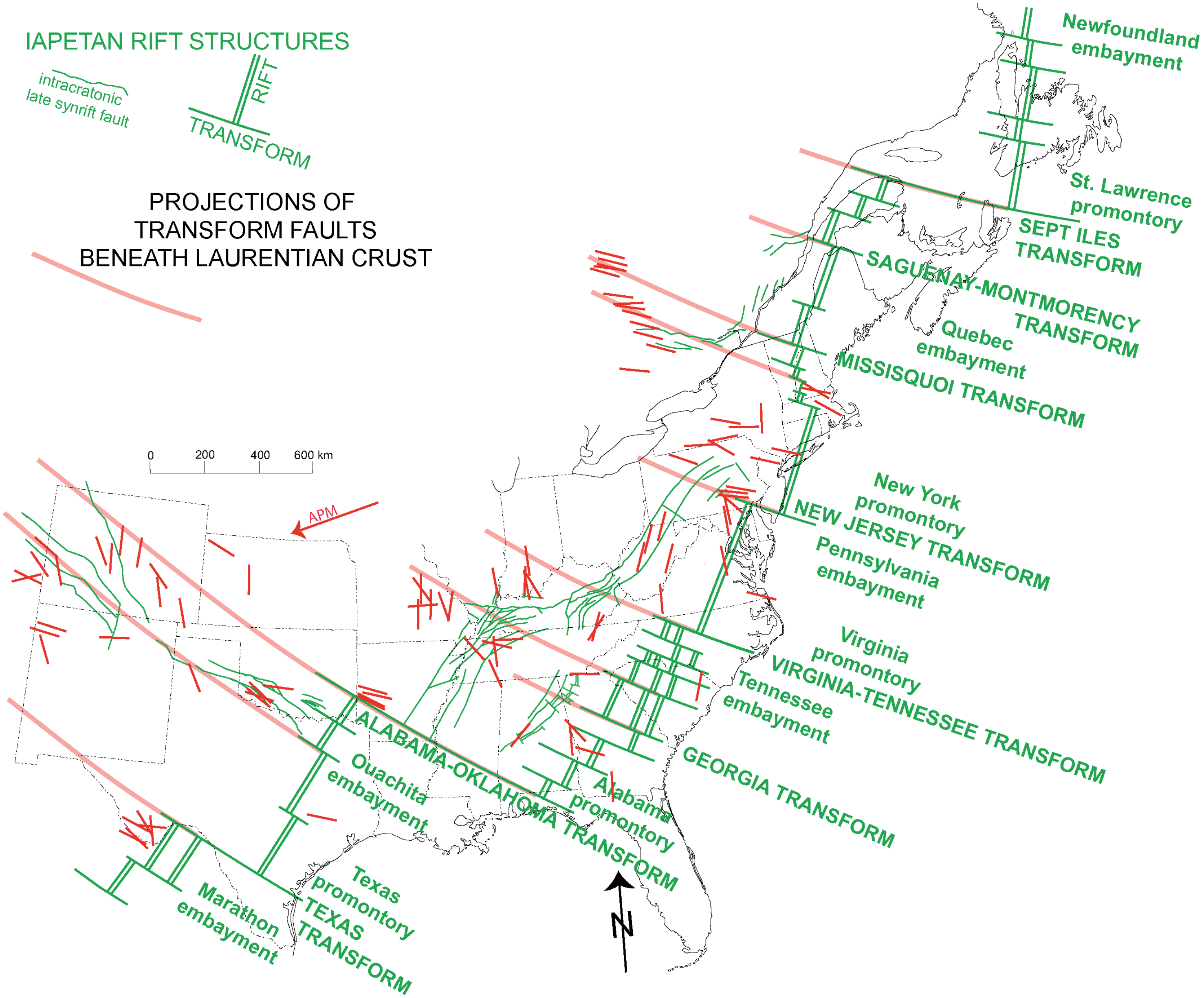

Figure 5. Map showing orientations of seismic anisotropy in relation to transform faults of the Iapetan rifted margin of Laurentia. Projections of transform faults (pink lines) into the continent are small circles extended to show inferred extent of transform-parallel distributed-shear fabric in the mantle lithosphere relative to transform-parallel fault systems in the crust. Fast directions of seismic anisotropy (short red lines) are filtered from the compiled map by Liu (2009), supplemented with ten additional points from Wagner et al. (2012), to show orientations that are >25° away from the direction of absolute plate motion (APM, shown by red arrow). Fast directions approximately parallel with absolute plate motion are interpreted to be from the asthenosphere; other orientations are likely from the mantle lithosphere.

suggest that the orientations of seismic anisotropy record zones of distributed-shear fabric along transform faults in the mantle lithosphere. Further work on seismic anisotropy is needed to better define the extent of transform-parallel fabrics in the mantle lithosphere. The fabrics in the mantle lithosphere are a key to tectonic inheritance of rift-related structures, specifically transform faults, during passive-margin thermal subsidence and synorogenic tectonic-load-driven flexural subsidence along the continental margin.

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