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Linking accretionary orogenesis with supercontinent assembly

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Abstract

Age relations for assembly of Gondwana and Pangea indicate that the timing of collisional orogenesis between amalgamating continental bodies was synchronous with subduction initiation and contractional orogenesis within accretionary orogens located along the margins of these supercontinents. Final assembly of Gondwana occurred between *c.*570 and 510 Ma, amalgamating the various components of East and West Gondwana. This was coeval with a switch from passive margin sedimentation to convergent margin activity along the Pacific margin of the supercontinent. Timing of subduction initiation along the Pacific margin ranges from 580 to 550 Ma as evidenced by the first appearance of arc derived detrital zircons in the upper Byrd Group sediments and the oldest supra-subduction zone plutons along the Antarctic segment of the margin. A phase of extension marked by supra-subduction zone ophiolite generation at 535–520 Ma is preserved in greenstone successions in eastern Australia and overlaps the onset of Ross–Delamerian contractional orogenesis between 520 and 490 Ma, inboard of the plate margin that coincides with the cessation of collisional orogenesis between the amalgamating blocks of Gondwana. Supra-subduction zone igneous activity was continuous throughout this period indicating that subduction was ongoing.

The final stages of assembly of the Pangean supercontinent occurred between *c.*320 and 250 Ma. Major plate boundary reorganization during this time was accompanied by regional orogenesis along the Pacific margin. The East Gondwana margin segment experienced transpressional and transtensional activity from *c.*305 Ma until *c.*270 Ma, after which convergence along the plate margin was re-established. In eastern Australia this involved a migration of arc magmatism eastward into the old subduction complex indicating a stepping out of the plate margin. Synchronous with this phase of plate re-adjustment was the Gondwanide Orogeny (305–230 Ma) affecting the entire Pacific margin of Pangea.

Temporal relations across supercontinents between interior collisional and marginal accretionary orogenies suggest a linked history between interior and exterior processes perhaps related to global plate kinematic adjustments. Orogenesis in accretionary orogens occurs in the absence of colliding bodies during ongoing subduction and plate convergence and must therefore be driven by a transitory coupling across the plate boundary. Correspondence of coupling with, or immediately following, subduction initiation and plate boundary reorganization, suggests it may reflect plate re-adjustments involving a temporary phase of increased relative convergence across the plate boundary.

Keywords: Accretionary Orogen; supercontinents; driving mechanisms; Gondwana; Pangea; plate reorganization

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1. Introduction

A fundamental question in tectonic studies of orogenic belts is *what drives orogenesis?* For classic 'collisional orogens' (Fig. 1a), where two continents have been brought together at the completion of a Wilson Cycle (e.g. Wilson, 1966), orogenesis reflects the resistance of a buoyant continental nuclei to subduction resulting in significant lithospheric thickening and deformation of both the upper and lower plates. However, in the case of *accretionary orogens* (Fig. 1b) the driving mechanism is less obvious because deformation, metamorphism and crustal growth take place in an environment of long-term subduction and plate convergence without the collision of continental blocks or large-scale buoyant lithosphere (Murphy and Nance, 1991; Windley, 1992; Sengör et al., 1993; Windley, 1993; Nance and Murphy, 1994; Sengör and Natal'in, 1996a,b). The lack of an obvious colliding body in accretionary orogens means that the driving mechanism of orogenesis must involve some form of transitory plate coupling which raises the question; *what mechanisms can increase coupling between plates within an accretionary orogenic system?*

The aim of this study is to establish the possible mechanisms of coupling by documenting the distribution and timing of orogenic events both across and laterally along an entire marginal orogen and compare these to orogenic events occurring in the same time period associated with amalgamation of cratonic blocks during supercontinent assembly. This will be achieved by reviewing and comparing the large volume of published geological, geochronological, geochemical and tectonic data of events related to the assembly of the supercontinents of Gondwana and Pangea with the Neoproterozoic to late Palaeozoic Terra Australis Orogen (Cawood, 2005) that was active along the Pacific margin of these supercontinents.

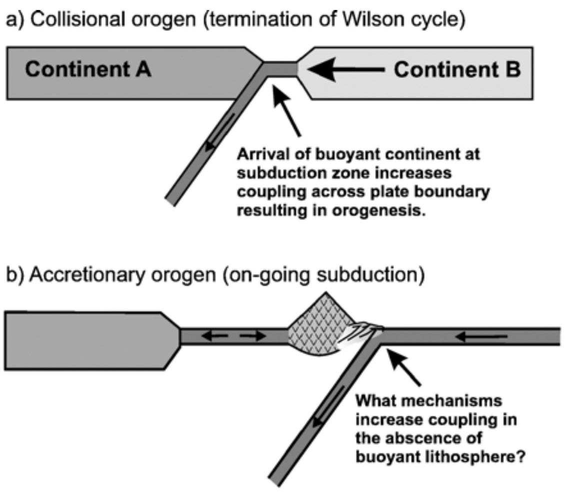

Fig. 1. Schematic diagram illustrating differences in tectonic plate interaction in collisional (a) and accretionary (b) orogens. In collisional orogens subduction of oceanic lithosphere will cease when the continents collide and crustal shortening of one or both continents takes place whereas, in accretionary orogens subduction is ongoing throughout orogenesis and shortening in the upper plate must be driven by coupling between the subducting and overriding plates.

This study demonstrates that the final stages of collisional orogenesis between the amalgamating blocks of the supercontinents were coeval with subduction initiation and accretionary orogenesis along the Pacific margin. Plate re-adjustments, caused by a change in relative plate convergence following cessation of shortening in the internal collisional orogens of the amalgamated supercontinents, provide a compelling means of driving increased transitory coupling across the plate boundary resulting in orogenesis.

2. Data selection and compilation

The tectonostratigraphic data presented in this study have been compiled from radiometric and palaeontological age data that are freely available in the published literature. The interpretation of the tectonic significance of the data was made using accompanying geochemical, structural and geological evidence. Emphasis was placed on distinguishing between an orogen and an orogeny; the former referring to the geographic extent of a tectonostratigraphic assemblage of rock units spanning an extended period of time that have been variably affected by one or more short-lived tectonothermal events, whereas an orogeny is a temporally specific tectonothermal event resulting in deformation, metamorphism and crustal thickening. The data have been sourced from a variety of isotopic techniques but, where available, U–Pb zircon data have been used for interpretation due to the more robust nature of this system that limits its potential for resetting during subsequent events. In those cases where less robust systems are the only data available, consideration of closure temperature and the potential for isotopic resetting by other means, such as deformation in the $^{40}Ar/^{39}Ar$ system, has been considered. The data are compiled in an un-filtered way in order to fully assess the level of coverage available.

Owing to the focussed nature of the current study, compilation has been restricted to two time intervals between 610 and 470 Ma, covering the time of Gondwana assembly, and 400 and 200 Ma, covering the time of Pangea assembly. For the period between 610 and 470 Ma some data for East Gondwana were sourced

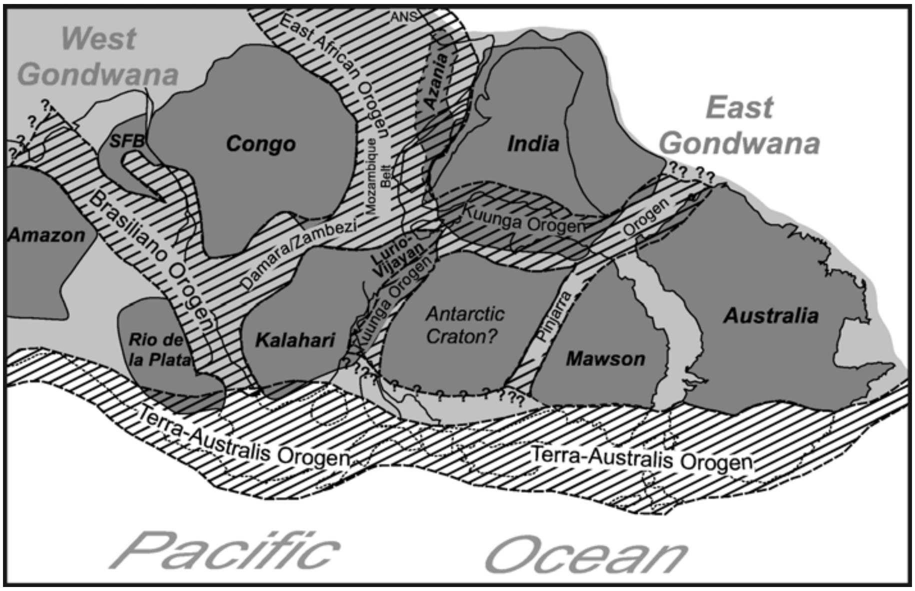

Fig. 2. Palaeozoic to Mesozoic reconstruction of Gondwana. Outlines of the main orogens formed during assembly of the component cratons of Gondwana are based on the following sources; East African Orogen and Mozambique Belt (Collins et al., 2003a; Johnson and Woldehaimanot, 2003; Meert, 2003; Collins and Pisarevsky, 2005; Fitzsimons and Hulscher, 2005), Damara/Zambezi Orogen (Goscombe et al., 2000; Hargrove et al., 2003; John et al., 2004; Johnson and Oliver, 2004), Brasiliano Orogen (Reid et al., 1991; Moura and Guadette, 1993; Trompette, 1997; Frimmel and Frank, 1998; Pedrosa-Soares et al., 2001), Kuunga Orogen (Meert and Van der Voo, 1997; Meert, 2003; Boger and Miller, 2004; Collins and Pisarevsky, 2005), and Pinjarra Orogen (Fitzsimons, 2003a,b). Inclusion of a joint Kalahari–Lurio–Vijayan Craton and Azania Cratonic block follows the new cratonic subdivision scheme of Collins and Pisarevsky (2005). Position of the Terra Australis orogen formed along the Pacific margin of Gondwana from the late Neoproterozoic to Mesozoic after Cawood (2005). Abbreviations: SFB — Sân Francisco block; ANS — Arabian Nubian Shield.

from the database of Meert (2003) that covers the age range 800 to 400 Ma[2]. The data have been plotted on time–space diagrams in order to illustrate the synchroneity of events across a variety of orogens located in both the internal and external portions of the supercontinents.

3. Gondwana related Events (610–470 Ma)

3.1. Gondwana internal assembly

The final assembly of Gondwana has generally been considered as a simple process involving the closure of the Mozambique Ocean separating East Gondwana (Australia, Antarctica, India, Madagascar and Arabia) from West Gondwana (Africa and South America). The final suture was thought to lie along the East African Orogen and Mozambique Belt and to have occurred at around 600 Ma (e.g. McWilliams, 1981; Stern, 1994). However, recently a number of studies have shown that East and West Gondwana may not have existed as independent and coherent continental masses in their own right and that the final amalgamation of Gondwana involved a more complex accretion of blocks along a variety of orogenic belts (Fig. 2) largely between around 570 and 520 Ma (Fitzsimons, 2000a,b; Collins et al., 2003a,b; Meert, 2003; van de Flierdt et al., 2003; Boger and Miller, 2004; Johnson and Oliver, 2004; Collins and Pisarevsky, 2005; Fitzsimons and Hulscher, 2005). These belts include the following; East African Orogen and Mozambique Belt lying between the Congo Craton and Arabian Nubian Shield, Azania and India; Kuunga Orogen separating western Antarctica, India and the Kalahari–Lurio–Vijayan Craton broadly following the definition of Meert (2003), except that the Zambezi–Damaran Orogen, between the Kalahari and Congo cratons, is considered separately; the Pinjarra Orogen between Australia–Mawson and India; and the Brasiliano Orogen (e.g. Moura and Guadette, 1993) between the Sao-Francisco–Rio de la Plata and Amazon cratons. Outlined below is a summary of key events related to craton assembly in each of these orogens.

3.1.1. East African Orogen and Mozambique Belt

The earliest events related to convergent terrane assembly in the East African Orogen occur in the Arabian Nubian Shield, which is composed of several

[2] Available online at (www.clas.ufl.edu/users/jmeert).

Neoproterozoic arc and ophiolite terranes (*c.* 870 to 650 Ma) and older continental fragments (Abdelsalam et al., 1998, 2002; Johnson and Woldehaimanot, 2003). The various terranes of the Arabian Nubian Shield were assembled in an intra-oceanic environment between *c.* 750 and 600 Ma, evidenced by syntectonic calc–alkaline granite plutons intruded on the boundaries of terranes (Fig. 3; Abdelsalam et al., 2002; Johnson and Woldehaimanot, 2003) before being accreted to the Saharan Metacraton by 580 Ma based on $^{40}Ar/^{39}Ar$ biotite and hornblende ages from a deformed granite in the Keraf Suture in northern Sudan that suggest rapid uplift and cooling at this time (Fig. 3; Abdelsalam et al., 1998; Abdelsalam et al., 2002).

In Tanzania, peak granulite metamorphism is identified between 655 and 610 Ma, (Coolen et al., 1982; Muhongo and Lenoir, 1994; Maboko and Nakamura, 1995; Möller et al., 2000; Muhongo et al., 2001; Kröner et al., 2003; Sommer et al., 2003) on the basis of U–Pb dating of metamorphic zircon and monazite within granulite facies gneisses. In southern Madagascar, de Wit et al. (2001) reported U–Pb monazite ages of 630–607 Ma within the Ampanihy and Vorokoftra shear zones, which they attributed to an early high pressure granulite event. These events correlate with similar aged granulite terranes in southern Kenya and northern Mozambique (Kröner et al., 1997; Meert and Van der Voo, 1997; Collins and Pisarevsky, 2005). In Ethiopia, Yibas et al. (2002) identified the Moyale phase of granite emplacement (700–550 Ma) that they associated with an active-margin setting during closure of the Moyale oceanic basin and amalgamation of the Arabian Nubian Shield. Yibas et al. (2002) further suggested that granites in the region intruded after 550 Ma were post-orogenic in nature, consistent with $^{40}Ar/^{39}Ar$ ages interpreted as recording uplift and cooling at the end of the East African Orogeny.

The events documented between 750 and 600 Ma appear to be confined to the northern sections of the East African Orogen from the Arabian Nubian Shield through Tanzania, Ethiopia, Somalia, Kenya, northern Mozambique and southern Madagascar, but are difficult to trace further south suggesting a diachroneity or discontinuity in the nature of orogenesis within the East African Orogen. In addition, many studies have suggested the existence of an easterly younging of deformation and metamorphic events within the orogen (Stern, 1994; Meert, 2003; Boger and Miller, 2004; Collins and Pisarevsky, 2005; Fitzsimons and Hulscher, 2005). The hypothesis of easterly younging is based on data suggesting two peaks of tectonic activity post-dating assembly of the Arabian Nubian Shield. These resulted in widespread high-grade metamorphism and granite intrusion throughout the Mozambique Belt between *c.* 580 and 530 Ma in Ethiopia, Malawi, central and western Madagascar and Sri-Lanka, and a younger event at *c.* 520 Ma concentrated in eastern Madagascar and SW India (Fig. 3; Holzl et al., 1994; Kroner et al., 1996; Ashwal et al., 1999; Kröner et al., 1999, 2000; de Wit et al., 2001; Fernandez et al., 2003; Cox et al., 2004; Collins and Pisarevsky, 2005; Fitzsimons and Hulscher, 2005; Fig. 3; Montel et al., 1994; Paquette et al., 1994; Montel et al., 1996; Paquette and Nedelec, 1998; Tucker et al., 1999; Kröner et al., 2001; Ring et al., 2002; Yibas et al., 2002; Meert, 2003). These two events are particularly well defined in Madagascar. Evidence for the older event includes the presence of 560 Ma metamorphic rims on detrital zircons of the Itremo and Molo groups (west-central Madagascar; Cox et al., 2004), granites within the Antananarivo block with intrusion ages of 560–530 Ma (Kröner et al., 2000), and widespread granulite grade metamorphism in charnokitic assemblages in many granites that are dated at *c.* 550 Ma (Kröner et al., 2000). Conversely in eastern Madagascar, kyanite schists of the Betsimisaraka Suture contain zircons with metamorphic rims dated at 518 ±7 Ma (Collins et al., 2003b) and monazites dated at 517±1 Ma (Fitzsimons et al., 2004) suggesting that deformation is younger in the east. These latter events are the youngest recorded high-grade metamorphic events in the East African Orogen and are suggested to mark final assembly of terranes within the orogen and collision of the Indian Craton with the amalgamated Tanzanian Craton, Saharan Metacraton and Arabian Nubian Shield.

Identification of three deformation pulses in the East African Orogen involving assembly of the Arabian Nubian Shield between *c.* 750 and 600 Ma, followed by high-grade metamorphism in the Mozambique Belt at *c.* 560 to 530 Ma (Fig. 3) and a final pulse at *c.* 520 Ma in the eastern Mozambique Belt, has led many workers (e.g. Meert, 2003; Collins and Pisarevsky, 2005; Fitzsimons and Hulscher, 2005) to question the model of Stern (1994) involving prolonged collisional assembly of East and West Gondwana continents along the East African Orogen. In addition, the localisation of the 750–600 Ma events within the northern segment of the orogen, coupled with the lack of contemporaneous events in India, and the widespread occurrence of younger collision-related events elsewhere in Gondwana, has led to the need for models that can account for this spatial variation and diachroneity of events. One such model suggests an initial collision between a newly defined microcontinental block named Azania, comprising Central Madagascar and parts of the Arabian Nubian Shield, with the Congo Craton at around *c.*570 to 500 Ma along the East African Orogen and then a subsequent second collision between India and Azania between *c.*530 and 510 Ma (Collins et al., 2003a,b; Collins and Pisarevsky, 2005). An alternative model by Fitzsimons and Hulscher (2005), suggests terrane transfer from the Congo to India prior to final closure of the Mozambique Ocean and prolonged collision along the East African Orogen.

3.1.2. Damara/Zambezi, Kuunga, and Pinjarra Orogens

The term Kuunga Orogen was first used by Meert et al. (1995) to describe the zone of collisional deformation associated with amalgamation of the Antarctic/Australian Craton with Indian and the Kalahari cratons between 550 and 530 Ma based on palaeomagnetic reconstructions. The orogen was later modified by Meert (2003), based on geochronological arguments, to include the area between the Congo and Kalahari cratons. In this study, the Kuunga Orogen as defined by Meert (2003) has been divided into three separate areas (Fig. 2) that are broadly coeval in terms of the timing of collisional activity, but are considered separately because they separate different cratonic blocks that may have been independent of each other prior to 520 Ma. The divisions chosen are, from west to east, the Damara/Zambezi Orogen (van de Flierdt et al., 2003; Johnson and Oliver, 2004) separating the Kalahari–Lurio–Vijayan and Congo cratons, the newly defined Kuunga Orogen separating the Antarctic craton from the Kalahari–Lurio–Vijayan and Indian cratons, and the Pinjarra Orogen separating the Australian–Mawson craton from the Antarctic and Indian cratons, as defined by Fitzsimons (2003b), but including the suggested extension through Lake Vostok (Figs. 2 and 4) to the Trans Antarctic Mountains (Fitzsimons, 2003a).

In models that consider West Gondwana as a coherent continental mass prior to Gondwana amalgamation (e.g. Stern, 1994), the Kalahari and Congo cratons were considered to have been amalgamated since at least the Mesoproterozoic largely based on the correlation of the Irumide Belt with the Chomo–Kalomo block south of the Zambezi Belt (Hanson et al., 1994; Wilson et al., 1997). However, recent geochronological studies have demonstrated that such a correlation is not valid (De Waele et al., 2003). In addition, the Mesoproterozoic Kaourera arc and Chewore ophiolite rocks of the Zambezi Belt (Johnson and Oliver, 1998, 2000, 2004) along with 595 Ma eclogites of Central Zambia (John et al., 2003, 2004) provide evidence that an ocean basin as well as an active subducting margin existed between the Kalahari and Congo during the Meso-Neoproterozoic indicating that they were still separated at this time (John et al., 2004; Johnson and Oliver, 2004). Based on new evidence, closure of the ocean separating the Congo and Kalahari cratons is now suggested to have occurred between 550 and 520 Ma (Fig. 3), resulting in high-pressure metamorphism in the Lufilian Arc (John et al., 2003, 2004) and Kadunguri whiteschists (Johnson and Oliver, 2002), thrust emplacement of layered orthogneises in northern Zimbabwe (Hargrove et al., 2003) and metamorphic overprinting of the Kaourerea Arc rocks (U–Pb 517±5 Ma rims on zircons Johnson and Oliver, 2004) and the Chewore Inliers (Goscombe et al., 2000). The final stages of this deformation are indicated by amphibolite facies retrogression of Archaean basement rocks in northeast Zimbabwe starting at about 510 Ma (Fig. 3; 507.9±2.5 Ma and 491.3±2.1 Ma, $^{40}Ar/^{39}Ar$ hornblende, Vinyu et al., 1999). Farther west in the Damara Belt (Fig. 2), a similar convergent history is recorded by high-grade metamorphism and migmatite generation between 540 and 510 Ma (Fig. 3) in the Oetmoed Granite–Migmatite Complex (Jung et al., 2000) and intrusion of associated quartz diorites in the Bandombaai Complex in Namibia (U–Pb 540±3 Ma van de Flierdt et al., 2003). These new data suggest that docking of the Kalahari and Congo cratons along the Damara/Zambezi Orogen was broadly synchronous with final suturing events along the East African Orogen (Fig. 3).

Along the southwest branch of the Kuunga Orogen (Fig. 4), high-grade metamorphism and collisional deformation are recorded in Dronning Maud Land at *c.* 570 to 550 Ma (Fig. 3) overprinting Meso-Neoproterozoic rocks under granulite conditions (Jacobs et al., 1998; Piazolo and Markl, 1999; Bauer et al., 2003; Jacobs et al., 2003a). A later phase of high-temperature metamorphism with associated granite generation occurred 20 m.y. later between 530 and 490 Ma (Fig. 3) and is suggested to have been caused by extensional collapse and tectonic escape (Jacobs et al., 2003d; Jacobs and Thomas, 2004). Along the northeastern branch of the Kuunga Orogen, between northern Antarctica and western India (Fig. 4), rocks in the Prydz Bay region record high-temperature metamorphism, migmatite and anatectic leucogneiss generation associated with compressional deformation between *c.* 535 and 525 Ma (Fig. 3; Carson et al., 1996; Fitzsimons et al., 1997; Fig. 3; Zhao et al., 1992; Henson and Zhou, 1995; Zhao et al., 2003). Coeval granite intrusion dated at 534±5 Ma (zircon cores), closely followed by high-temperature metamorphism of the granites at 529±14 Ma (zircon rims), is recorded in the Grove Mountains east of Prydz Bay (Zhao et al., 2003).

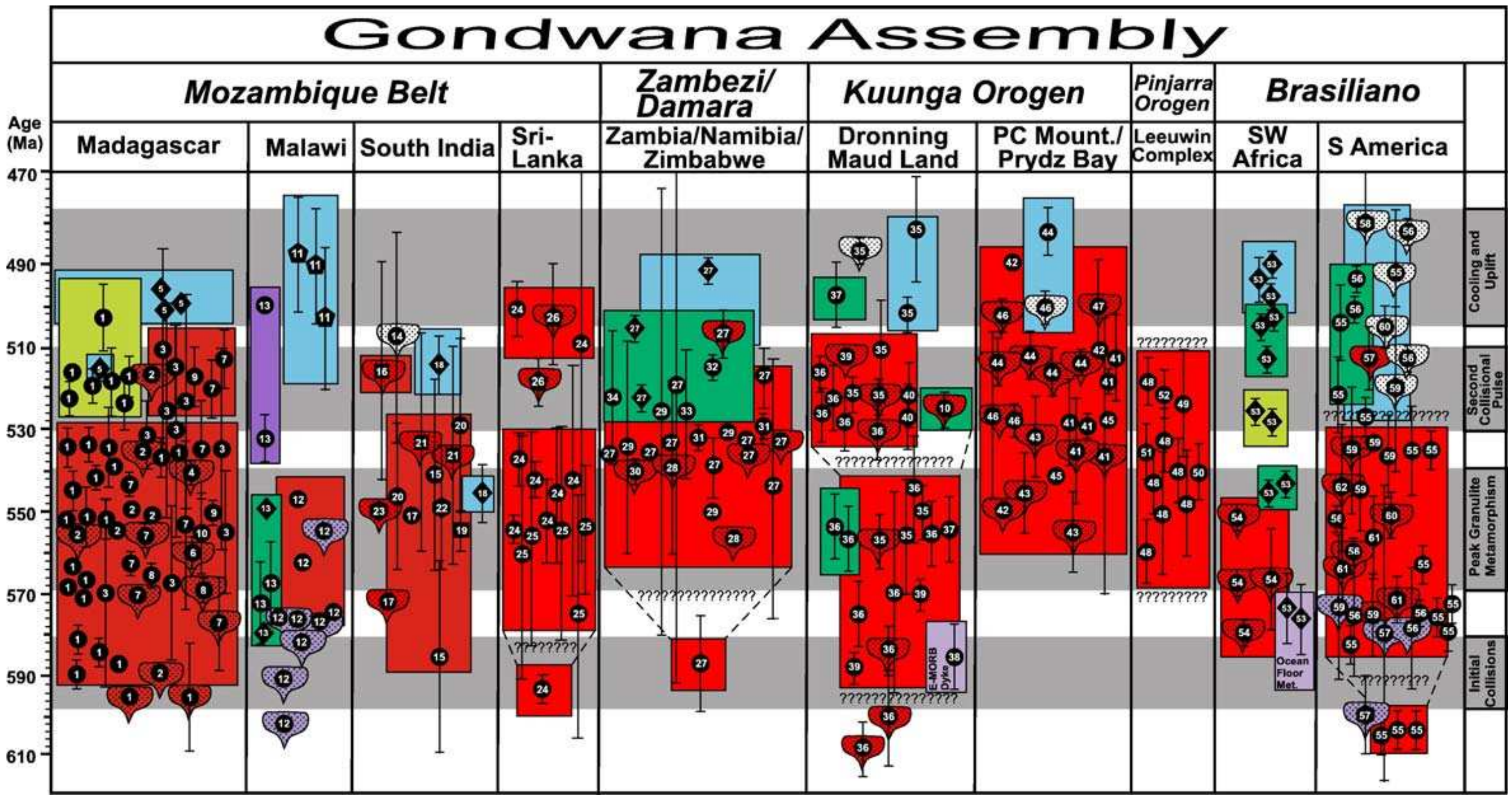

Fig. 3. Compilation of tectonostratigraphic data associated with assembly of Gondwana. The data indicate a correlation of events across the various orogens with initial collision of cratonic blocks occurring at around 590 Ma, followed by a thermal peak resulting in widespread granulite grade metamorphism and magmatism associated with crustal shortening between 570–550 Ma, a second phase of high-grade metamorphism and crustal shortening around 520 Ma and final cooling and uplift by around 490 Ma. Numbers on data points refer to the following sources: 1 (de Wit et al., 2001), 2 (Kröner et al., 2000), 3 (Montel et al., 1994), 4 (Tucker et al., 1999), 5 (Meert et al., 2003), 6 (Paquette and Nedelec, 1998), 7 (Paquette et al., 1994), 8 (Kröner et al., 1999), 10 (Cox et al., 2004), 11 (Cahen and Snelling, 1966), 12 (Kröner et al., 2001), 13 (Ring et al., 2002), 14 (Berger and Braun, 1997), 15 (Braun et al., 1998), 16 (Unnikrishnan-Warrier et al., 1995), 17 (Kovach et al., 1998), 18 (Rathore et al., 1999), 19 (Choudhary et al., 1992), 20 (Bartlett et al., 1998), 21 (Unnikrishnan-Warrier, 1997), 22 (Hansen et al., 1985), 23 (Santosh and Drury, 1988), 24 (Holzl et al., 1994), 25 (Kröner et al., 1994), 26 (Fernando and Izumi, 2001), 27 (Vinyu et al., 1999), 28 (Müller et al., 2000), 29 (Hargrove et al., 2003), 30 (van de Flierdt et al., 2003), 31 (John et al., 2004), 32 (Johnson and Oliver, 2004), 33 (Goscombe et al., 2000), 34 (Goscombe et al., 1998), 35 (Jacobs et al., 2003a), 36 (Jacobs et al., 1998), 37 (Jacobs et al., 2003c), 38 (Bauer et al., 2003), 39 (Mikhalsky et al., 1995), 40 (Jacobs et al., 2003b), 41 (Fitzsimons et al., 1997), 42 (Boger et al., 2001), 43 (Zhao et al., 1992), 44 (Carson et al., 1996), 45 (Zhang et al., 1996), 46 (Zhao et al., 2003), 47 (Kinny et al., 1993), 48 (Janssen pers. comm. 2005), 49 (Nelson, 1996), 50 (Nelson, 1999), 51 (Nelson, 2002), 52 (Collins, 2003), 53 (Frimmel and Frank, 1998), 54 (Seth et al., 1998), 55 (Machado et al., 1996), 56 (Heilbron and Machado, 2003), 57 (Söllner et al., 1987), 58 (Besang et al., 1977), 59 (Söllner et al., 1991b), 60 (Siga, 1986), 61 (Söllner et al., 1989), 62 (Tassinari, 1988).

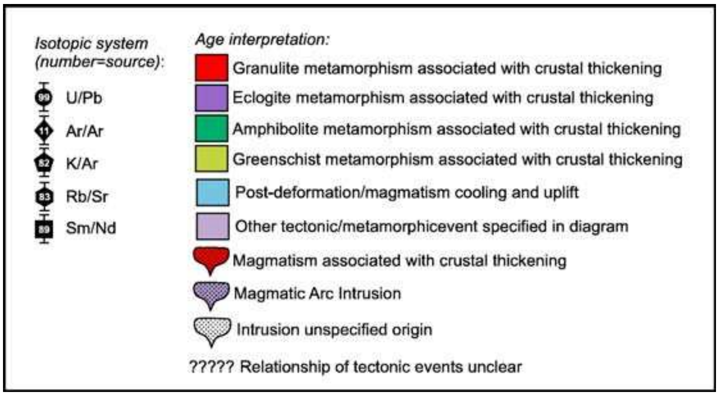

Fig. 3 (*continued*).

Archean layered igneous complexes, within the Rauer Group, Prydz Bay (Fig. 4) are overprinted by high-temperature deformation at 519±8 Ma (Fig. 3; Harley et al., 1998). Deformation on the Indian side of the Kuunga Orogen is poorly constrained, but metamorphism associated with compressional deformation and shear zone reactivation is documented in the Eastern Ghats of southern India between 550 and 500 Ma (Mezger and Cosca, 1999; Dobmeier and Raith, 2003).

Data from the Pinjarra Orogen (Figs. 3 and 4) are limited and, due to exposure restrictions caused by the Antarctic ice cap, are only available from the Australian section. However, the data are of high quality and indicate granite intrusion between *c.* 540 and 525 Ma (Fig. 3; Janssen pers. comm. 2005; Nelson, 1996, 1999, 2002; Collins, 2003) and granulite–amphibolite grade metamorphism and migmatite generation between 550 and 520 Ma (Janssen pers. comm. 2005; Collins, 2003). The coeval history of the Pinjarra and Kuunga orogens has led some authors to propose that they may be linked suggesting that either the Mawson and Antarctic cratons are in fact one body or that if an orogen separates them, it is unrelated to the Pinjarra orogen (Fitzsimons, 2003b; Meert, 2003; Veevers, 2004). However, Fitzsimons (2003a) identified structural features separating the Mawson and Antarctic cratons beneath the Antarctic ice cap and noted similarities between features in the Transantarctic Mountains and the Darling ranges of Western Australia. Based on this evidence, we extend the Pinjarra orogen through Antarctica dividing it into the Mawson and Antarctic cratons (Fig. 4).

3.1.3. Brasiliano Orogen

The Brasiliano Orogen records deformational, magmatic and metamorphic events associated with the closure of the Adamaster Ocean and amalgamation of the combined Amazon and Rio de la Plata cratons with the Congo and Kalahari Cratons. Rift-related volcanics (c. 900 Ma, Machado et al., 1989) and inferred ophiolite remnants (*c.* 800 Ma) preserved in the Araçuaí–West-Congo orogen (Fig. 5) provide evidence for a Neoproterozoic ocean basin in the northern section of the Brasiliano Orogen (Pedrosa-Soares et al., 2001). Closure of this ocean resulted in development of magmatic arc related granites at *c* 620–570 Ma (Weidmann, 1993; Campos-Neto and Figueiredo, 1995; Pedrosa-Soares et al., 1999), on the Brazilian side of the orogen and further south into the Ribeira Belt (Fig. 5; Heilbron and Machado, 2003; Fig. 5; Machado et al., 1996). These overlap with *c.* 590 to 570 Ma granites interpreted as collision related (Fig. 3; Pedrosa-Soares et al., 2001; Fig. 3; Weidmann, 1993) and with granulite facies metamorphism in southernmost Brazil (Fig. 3; Leite et al., 2000). A later suite of peraluminous granites known as the Almenara suite is locally intruded along the eastern side of the Araçuaí and is interpreted to mark the end of this collisional pulse and granulite metamorphism in this part of the orogen at *c.* 580 to 560 Ma (Fig. 3; Siga, 1986; Söllner et al., 1989). A final pulse of magmatism between 535 and 500 Ma (Söllner et al., 1991b; Weidmann, 1993; Machado et al., 1996; Pedrosa-Soares et al., 1999; Noce et al., 2000; Heilbron and Machado, 2003) is associated with charno–enderbitic facies granites and granodiorites intruded during a phase of strike–slip deformation, and is suggested to mark the final amalgamation of continental blocks (Weidmann, 1993; Pedrosa-Soares et al., 1999; Pedrosa-Soares et al., 2001; Schmidtt et al., 2004). A similar magmatic and metamorphic history is recorded in the Brasília belt of central Brazil, except that 590–500 Ma magmatism is here related to post-collisional uplift and extension of the Paraná Block and associated sedimentation in the Paraná Basin (Trompette, 1997; Pimentel et al., 1999).

The Brasiliano Orogen extends into South Africa along the Gariep and Kaoko Belts (Fig. 5). The Gariep belt records rift-related and oceanic magmatism around 900–750 Ma (Kröner, 1975; Allsop et al., 1979; Reid et al., 1991; Gresse and Scheepers, 1993; Frimmel and Frank, 1998). Closure of this ocean and subsequent continental collision resulted in granite magmatism and amphibolite to granulite facies metamorphism throughout the Gariep Belt between *c.* 570 and 540 Ma (Gresse and Scheepers, 1993; Frimmel and Frank, 1998; Frimmel, 2000), and a later lower grade amphibolite to greenschist facies event at *c.* 530–500 Ma (Fig. 3; Frimmel and Frank, 1998; Frimmel, 2000). Contemporaneous foreland basin sedimentation is recorded in the Nama and Varhynsdorp Groups and is interpreted to be sourced from the exhumed Gariep Belt rocks (Gresse and Scheepers, 1993; Gresse, 1995; Hälbich and Alchin, 1995). The Kaoko Belt lies to the north of the Gariep Belt and is though to be correlative with the West Congo Belt in Africa and Dom Feliciano and Riberia Belts of Brazil (Kröner and Correia, 1980; Trompette, 1997). The magmatic history of the Kaoko Belt is more restricted than that of the Gariep and records collision-related magmatism and high-grade metamorphism between about 650 and 550 Ma (Fig. 3; Heilbron and Machado, 2003; Fig. 3; Seth et al., 1998). A subsequent lower grade greenschist facies event associated with transpressional shearing along the belt is recorded by K–Ar ages of pelitic schists at *c.* 550–530 Ma (Ahrendt et al., 1983) that is related to cooling and uplift after the main collisional phase (Dürr and Dingeldey, 1996).

3.2. Gondwana Pacific margin

Neoproterozoic rifting of Laurentia from Rodinia resulted in the creation of the Iapetus and Pacific Oceans (Cawood et al., 2001; Cawood, 2005). Whilst the Iapetus ocean followed a Wilson cycle evolution of ocean closure and continental collision resulting in the Appalachian–Caledonian Orogen (Wilson, 1966), the Pacific has never closed and has been bounded by the margins of West Laurentia and Gondwana throughout its life (Dalziel, 1991; Hoffman, 1991; Moores, 1991; Coney, 1992). The Gondwana margin segment extending some 18,000 km from eastern Australia through New Zealand, Antarctica and South Africa to the southwest American coastline (Fig. 2) provides a remarkable record of initiation of the Pacific and its subsequent

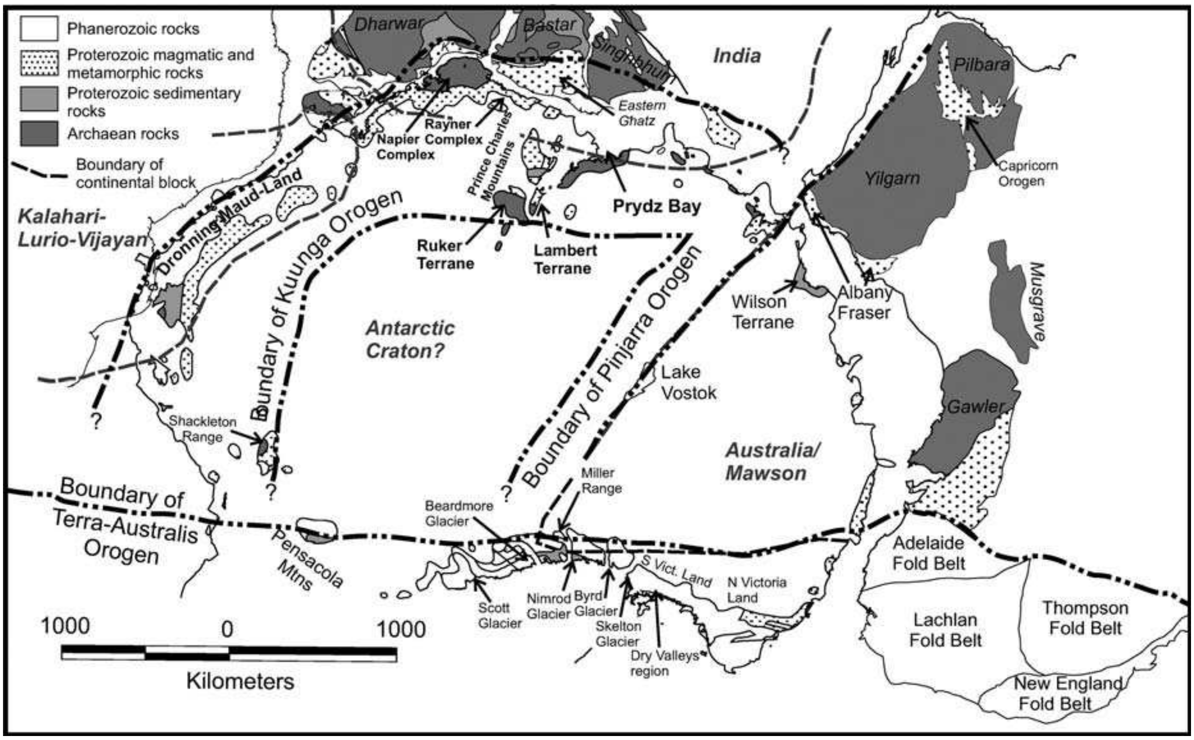

Fig. 4. Palaeozoic to Mesozoic reconstruction of south east Gondwana showing positions of the Kuunga (after Meert, 2003; Boger and Miller, 2004; Collins and Pisarevsky, 2005; after Meert and Van der Voo, 1997), Pinjarra (after Fitzsimons, 2003a,b), and Terra Australis Orogens (after Cawood, 2005). Geological provinces after Collins and Pisarevsky (2005).

history. The coeval history of this margin prior to the breakup of Gondwana in the late Mesozoic led Cawood (2005) to name this region the Terra Australis Orogen (Fig. 2). Whilst East and West Gondwana may not have existed as continental masses in their own right, for the purposes of this discussion they will be used as geographic subdivisions in which to present the data.

3.2.1. East Gondwana margin

The Pacific margin of East Gondwana (Fig. 4) extends from the eastern Australian mainland, through Tasmania and New Zealand, to the Transantarctic Mountains and Antarctic Peninsula (Stern, 1994; Cawood, 2005). Rift and passive margin successions related to Rodinia breakup are preserved in the thick Neoproterozoic to early Palaeozoic siliciclastic and carbonate successions preserved in the Adelaide Fold Belt, western New South Wales (NSW) and western Tasmania, the Anakie inlier of Queensland and the Transantarctic Mountains. The most complete record of the marginal sequence is preserved in the Adelaide fold belt where it accumulated in a series of rift and sag basins between *c.* 830 and 500 Ma when sedimentation was terminated during the Ross–Delamerian Orogeny (Preiss, 1987; Drexel et al., 1993; Powell et al., 1994; Drexel and Preiss, 1995; Preiss, 2000). In the Transantarctic Mountains, rifting and deposition of the Beardmore Group (as redefined by Goodge et al., 2004b) had occurred by at least 668 Ma, based on U–Pb ages of zircon from an interlayered, rift-related gabbro (Goodge et al., 2002, 2004b), and coeval accumulation of coarse siliciclastic rocks and limestones is evident in the Princes Anne Glacier region (Goodge et al., 2004b). Detrital zircon studies of early Palaeozoic sedimentary successions of the upper Bird group of the Transantarctic Mountains reveal that sandstones of the Starshot Formation contain fresh, locally derived igneous grains with ages in the range 580 to 520 Ma (Goodge et al., 2002; Myrow et al., 2002; Goodge et al., 2004a,b). These grains are interpreted to have been sourced from voluminous, continental arc magmatism suggesting a change from passive- to active-margin characteristics and showing subduction was active in the Antarctic segment of the margin by 580 Ma (Goodge et al., 2004a,

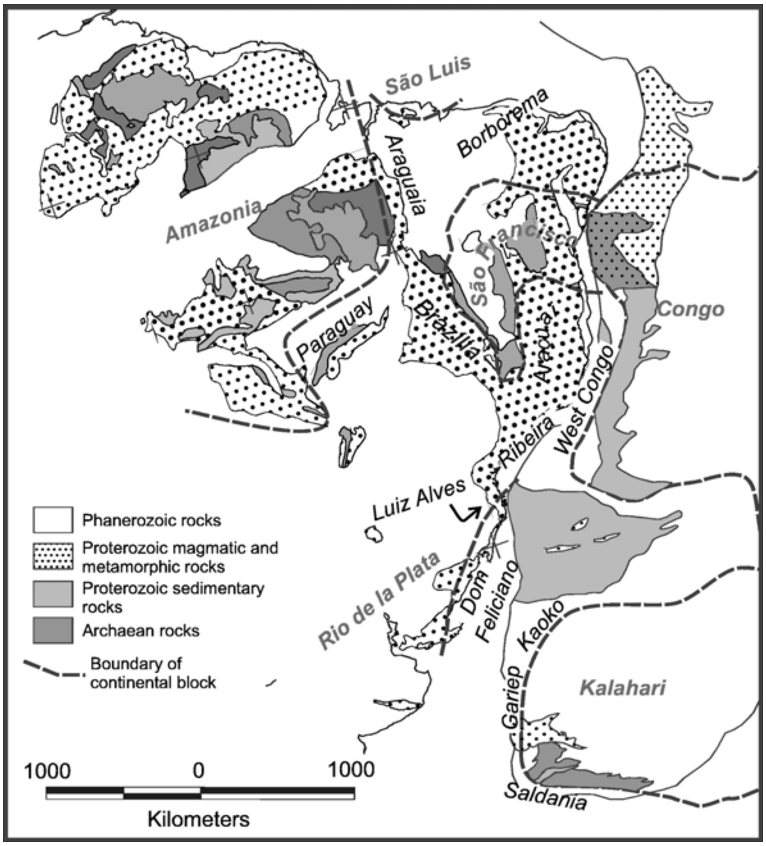

Fig. 5. Reconstruction of the Brasiliano Orogen *c.* 550–500 Ma showing positions of various tectonic belts and geological subdivisions (after Reid et al., 1991; Moura and Guadette, 1993; Frimmel and Frank, 1998; Pedrosa-Soares et al., 2001; Collins and Pisarevsky, 2005; after Trompette, 1997).

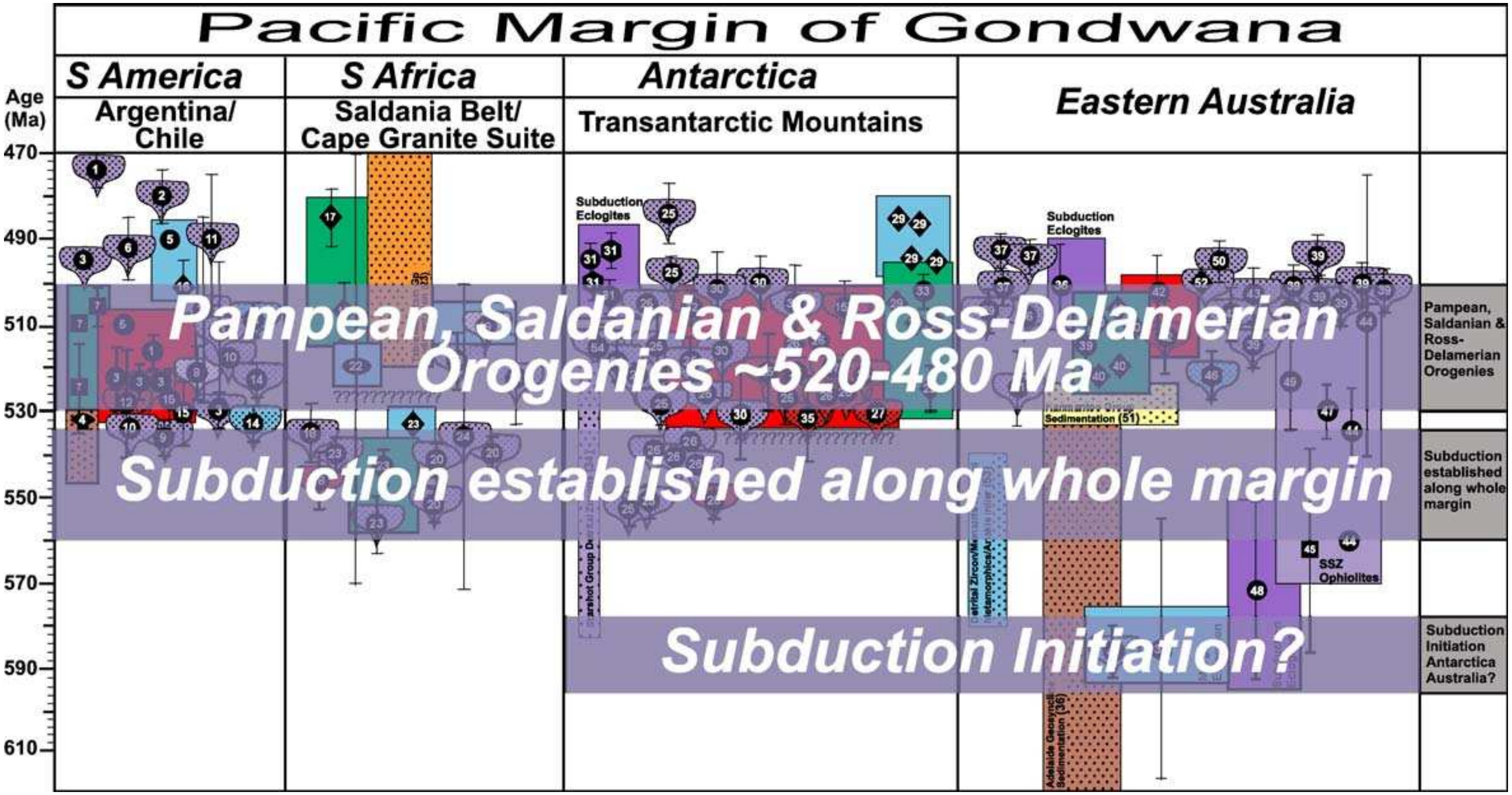

Fig. 6. Compilation of tectonostratigraphic data associated with tectonic events along the Pacific margin of Gondwana between 610–470 Ma. The data indicate a correlation of events along the margin with initiation of subduction in the East Gondwana segment at *c.* 590 Ma and *c.* 560–530 Ma in West Gondwana followed by margin compression and resultant orogenesis along the whole margin between around 530–500 Ma. Numbers on data points refer to the following sources: 1 (Gromet and Simpson, 1999), 2 (Pankhurst and Rapela, 1998), 3 (Rapela et al., 1998b), 4 (Durand, 1996), 5 (Fantini et al., 1998), 6 (Pankhurst et al., 1998a), 7 (Lucassen et al., 2000), 8 (Pankhurst et al., 2003), 9 (Omarini et al., 1999), 10 (Bachmann et al., 1987), 11 (von Gosen et al., 2002), 12 (Söllner et al., 2000), 13 (Sato et al., 1999), 14 (Rapela et al., 2003), 15 (Sims et al., 1998), 16 (Krol and Simpson, 1999), 17 (Gresse et al., 1992), 18 (Da Silva et al., 2000), 19 (Chemale et al., 1998), 20 (Scheepers and Armstrong, 2002), 21 (Schoch and Burger, 1976), 22 (Jacobs and Thomas, 2001), 23 (Armstrong et al., 1998), 24 (Dunlevey, 1981), 25 (Encarnación and Grunow, 1996), 26 (Goodge et al., 1993), 27 (Parkinson, 1994), 28 (Rowell et al., 1993), 29 (Goodge and Dallmeyer, 1992), 30 (Allibone and Wysoczanski, 2002), 31 (Vincenzo et al., 1997), 32 (Pankhurst et al., 1998b), 33 (Wysoczanski and Allibone, 2004), 34 (Goodge et al., 2004b), 35 (Cooper et al., 1997), 36 (Turner et al., 1998), 37 (Perkins and Walshe, 1993), 38 (Foden et al., 2001), 39 (Black et al., 1997), 40 (Foster et al., 2005b), 41 (Crawford et al., 1997), 42 (Foden et al., 1999), 43 (Foden et al., 2002), 44 (Aitchison and Ireland, 1995), 45 (Bruce et al., 2000), 46 (Turner, 1996), 47 (Aitchison et al., 1992), 48 (Watanabe et al., 1998), 49 (Maher et al., 1997), 50 (Crawford et al., 1996), 51 (Preiss, 1995), 52 (Foster et al., 1998), 53 (Fergusson et al., 2001), 54 (Van Schmus et al., 1997).

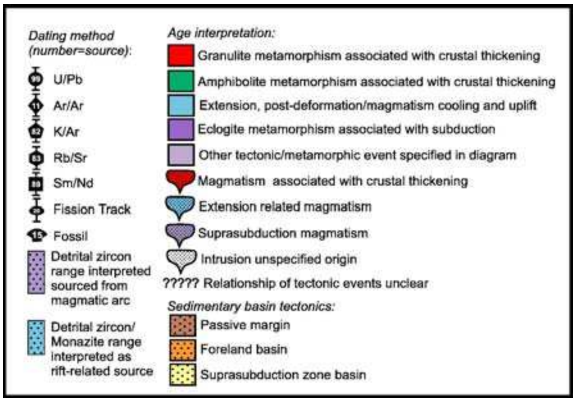

Fig. 6 (*continued*).

b, 2002). There are no definitive arc related intrusions in the Transantarctic Mountains and Victoria Land until *c.* 550 to 525 Ma (Fig. 6) when the Granite Harbour intrusives (Encarnación and Grunow, 1996), Liv Group bimodal suite (Wareham et al., 2001) and plutons of the Dry Valleys region (Allibone and Wysoczanski, 2002) were emplaced and calc–alkaline granites were intruded into the high-grade metamorphic basement gneisses of the Nimrod Group (Goodge et al., 1993). In the Queen Maud Mountains and Skelton Glacier, calc–alkaline igneous rocks may be as old as 550 Ma, representing possibly the earliest preserved arc intrusives along this part of the margin. However, the data are upper intercept ages on a discordia line and may represent complex inheritance and Pb loss (Rowell et al., 1993; Encarnación and Grunow, 1996; Van Schmus et al., 1997).

At around the same time as the transition from passive to active-margin tectonics along the Antarctic margin, the eastern Australian margin was reactivated during a phase of renewed extension and rifting. This resulted in the generation of the Mount Wright Volcanics in western NSW, where a transitional alkaline basalt–trachybasalt–trachyandesite–trachyte–alkali–rhyolite suite was erupted at 586±7 Ma (U–Pb zircon) and was subsequently overlain and intruded by a calc–alkaline basalt–andesite–dacite suite (Crawford et al., 1997). Margin extension is also suggested by the deposition and metamorphism of the early Palaeozoic units in the Anakie Inlier and Charters Tower Province of eastern Queensland. Detrital zircons from these rocks give ages ranging from *c.* 3100 to 500 Ma with the majority of grains lying between 600 and 500 Ma (Fergusson et al., 2001, 2007). The Archaean to Mesoproterozoic ages are interpreted to represent sources in the cratonic basement to the west, whereas the zircons (and monazites) in the younger 600–500 Ma range are suggested to at least incorporate some margin rift-related sources coeval with the magmatism in the Mount Wright volcanics (Fergusson et al., 2001, 2007). The onset of margin extension can be traced further south through the Adelaide Fold Belt where shales and basalts from below the Wonoka Formation define a 586±30 Ma Rb–Sr isochron. Sr-isotopic compositions of associated carbonate units indicate a phase of intra-basinal fluid flow thought to have been triggered by the onset of margin extension (Foden et al., 2001). In Tasmania, rift-related picrites and tholeiitic basalts on King Island have recently been dated at 579 Ma (Sm–Nd isochron Meffre et al., 2004) and may also be related to margin extension. It is interesting to note that unlike the Australian section of the margin there is no evidence in Antarctica for significant margin extension at the onset of subduction, suggesting a possible difference in the geometry of the subducting margin between the two areas. In Antarctica, subduction appears to have been initiated in the late Neoproterozoic close to the continent–ocean boundary with development of a continental margin arc but in eastern Australia, passive margin sedimentation continued, overlapping with subduction along the Antarctic segment, and subduction probably initiated outboard of the margin in an intra-oceanic setting (Fig. 7, Cawood, 2005).

The earliest occurrence of subduction related metamorphism along the East Gondwana margin segment is recorded by eclogites in the Peel Fault system of the New England Fold Belt, eastern Australia (Figs. 4 and 6). The eclogites are preserved within serpentinite melange and have a $^{206}Pb/^{238}U$ age of 571±22 Ma interpreted to reflect the age of peak metamorphism (Watanabe et al., 1998). The New England Fold belt also contains several late Neoproterozoic–early Palaeozoic supra-subduction zone ophiolites that are interpreted to

Fig. 7. Reconstruction of Pangea at 250 Ma showing positions of the Alleghanian–Ouachita, Varscan, Urals and Terra Australis orogens. Plate positions were reconstructed by S. A. Pizarevsky using Plates software produced by University of Texas at Austin.

have occupied a position outboard of the continental margin in a marginal basin or proto-arc position (Aitchison et al., 1992; Aitchison and Ireland, 1995; Bruce et al., 2000). The oldest of these is the Malborough ophiolite of the northern New England Fold Belt in Queensland, which has a crystallisation age of 562±22 Ma (Fig. 6; Sm–Nd isochron Bruce et al., 2000). In the Peel-Manning fault system of the southern New England Fold Belt ophiolites contain plagiogranites which have yielded U–Pb zircon ages of 530±6 Ma, 535±10 Ma and 509±30 Ma (Aitchison et al., 1992; Aitchison and Ireland, 1995). Supra-subduction ophiolites are also documented in Tasmania containing boninitic mafic–ultramafic sequences suggestive of a forearc/proto-arc environment of formation (Crawford and Berry, 1992). The only estimate of the crystallisation age of the ophiolites is provided by zircons separated from a late stage tonalite within the Heazlewood River Complex dated at 514±5 Ma (Black et al., 1997), which is close to estimates of the timing of their emplacement onto the passive margin sequences at 515–510 Ma (Crawford and Berry, 1992; Meffre et al., 2000; Crawford et al., 2003). In the western Victorian section of the Adelaide Fold Belt, unexposed ophiolites of the Dimboola Subzone discovered during borehole drilling are dated at 524±9 Ma and are thought to be along strike extensions of the Tasmanian ophioilites (U–Pb zircon; Maher et al., 1997). Calc–alkaline volcanics in the Mount Staveley Belt (500–495 Ma) provide evidence of supra-subduction magmatism at this time that may be associated with the Mount Read volcanic rocks of Tasmania (495–503 Ma) and the Delamerian and Tasmanian ophiolites (Perkins and Walshe, 1993; Crawford et al., 1996; Foster et al., 1998). The Lachlan Fold belt of eastern Australia contains several Cambrian age (*c.* 500 Ma) ophiolite bodies that occupied an intra-oceanic supra-subduction position before being imbricated with turbidite-dominated cover sequences during the late Ordovician to Devonian (Spaggiari et al., 2003). Along the Antarctic section of the margin there are no documented occurrences of ophiolite sequences *senso stricto* and the only occurrences of oceanic rocks are *c.* 500 Ma eclogite facies rocks (Fig. 6), interpreted as subducted oceanic crust, preserved in the Lanterman Range of Northern Victoria Land (Fig. 4; Peacock and Goodge, 1995; Capponi et al., 1997; Fig. 4; Ricci et al., 1996, 1997; Vincenzo et al., 1997). The lack of ophiolite rocks may reflect the lack of evidence of supra-subduction marginal basin extension within this section of the margin and the predominance of continental margin arc intrusions.

Following subduction initiation, the margin of East Gondwana experienced compressional deformation and mountain building during the Ross–Delamerian Orogeny. Estimates of the onset of this orogenic pulse vary along the margin and there is evidence that parts of the Antarctic segment may have been affected by compressional deformation earlier than the Australian segment. Ductile fabrics within calc–alkaline plutons in the Transantarctic Mountains suggest deformation at around 530 Ma (Encarnación and Grunow, 1996; Allibone and Wysoczanski, 2002) and may even have been developed as early as 550 to 540 Ma based on metamorphic rims of zircons in the Skelton Group (Wysoczanski and Allibone, 2004) and intrusion of plastically deformed granites within the Nimrod Group (Goodge et al., 1993). However, the main pulse of Ross–Delamerian orogenesis took place along the whole margin at *c.* 520 to 490 Ma (Foden et al., 2006). This main pulse was immediately preceded by basin formation and deposition of the Kanmantoo Group in the Adelaide Fold Belt between *c.* 532 and 526 Ma (Fig. 5; Flöttmann et al., 1994; Foster et al., 2005b; Gray and Foster, 2005; Fig. 5; Preiss, 1995). Evidence of peak metamorphism can be found in the Forth metamorphic complex of Tasmania, which preserve zircon metamorphic rims dated at 514±4.6 Ma (Black et al., 1997) and 511±4 Ma, suggesting peak metamorphic temperatures of 700–740 °C (Meffre et al., 2000). The earliest phase of deformation in the Adelaide Fold Belt is dated by the emplacement of the Rathjen Gneiss igneous precursor at 514±5 Ma, which was metamorphosed at 503±7 Ma based on metamorphic zircon rims (Fig. 5; U–Pb zircon Foden et al., 1999). Major changes in sedimentation are also recorded by the upper Byrd Group of the Transantarctic Mountains, which terminated the Shackleton carbonate platform succession with an influx of mud and silt deposits and the onlap of coarse alluvial–fluvial debris at around 515 Ma (Myrow et al., 2002; Goodge et al., 2004b). Detrital zircon and muscovite populations with an active-margin source in the upper Byrd Group suggest rapid denudation of recently formed continental arc basement during the Ross–Delamerian Orogeny (Myrow et al., 2002; Goodge et al., 2004a,b). Emplacement of the Tasmanian ophiolites at 515–510 Ma was caused by compression of the continental margin as a result of the Ross–Delamerian Orogeny (Crawford and Berry, 1992; Crawford et al., 2003; Boger and Miller, 2004; Cawood, 2005; Foster et al., 2005b; Gray and Foster, 2005). Subduction continued throughout the Ross–Delamerian orogenesis (Fig. 6) with intrusion of the calc–alkaline Granite Harbour suite in the Transantarctic Mountains (Encarnación and Grunow, 1996) and I- and S-type granites within the Delamerian Fold Belt (Foden et al., 2002; Turner et al., 1996) and Cambrian arc intrusions in New Zealand (Münker and Cooper, 1995; Münker, 2000; Münker and Crawford, 2000). Cessation of compressional Ross–Delamerian deformation and the onset of margin extension is indicated by the Mount Read and Mount Stavely volcanic complexes that are interpreted to have been intruded in rift environments above a west-dipping subduction zone between 505 and 495 Ma in Tasmania and eastern Australia (Fig. 5; Foster et al., 2005b).

3.2.2. West Gondwana margin

The West Gondwana margin segment of the Terra Australis Orogen extends through South Africa to South America (Fig. 2). Events here are less well constrained than their eastern counterparts largely due to pervasive overprinting by Permo-Triassic and younger events.

As was the case in the East Gondwana margin segment, the continental margin of West Gondwana evolved into a passive margin setting with siliciclastic and carbonate platform development following the breakout of Laurentia from Rodinia. The first indications of a transformation to an active margin and the onset of subduction are found in Argentina where metaluminous calc–alkaline granites and dacite–rhyolite bodies with supra-subduction chemical signatures were intruded into the Puncoviscana passive margin sequences in the Sierras de Córdoba at 530±4 Ma (Fig. 6, U–Pb zircon; Rapela et al., 1998b). The magmatic arc was short lived with crustal thickening and granulite grade metamorphism and deformation occurring at 522±8 Ma (Fig. 6, U–Pb monazite; Rapela et al., 1998a). This deformation was accompanied by generation of migmatites and granite intrusion dated at 523±2 Ma (Fig. 6, U–Pb zircon), which are suggested to have been associated with the closure of the Puncoviscan Ocean in the Early Cambrian (Rapela et al., 1998a). The creation and closure of the Puncoviscan Ocean are attributed to the accretion of the previously rifted Pampean terrane back onto the Gondwanan margin (Rapela et al., 1998b). Following accretion of the Pampean terrane there was a hiatus of tectonic activity between 515 and 500 Ma, after which the Famatanian magmatic arc was developed indicated by emplacement of high-Al trondjemites at 496±2 Ma in the Pampean foreland. The main arc had developed to the west by 490 Ma and was active until *c.* 460 Ma (Pankhurst and Rapela, 1998; Rapela et al., 1998a,b; Ramos, 2000; Pankhurst et al., 2001).

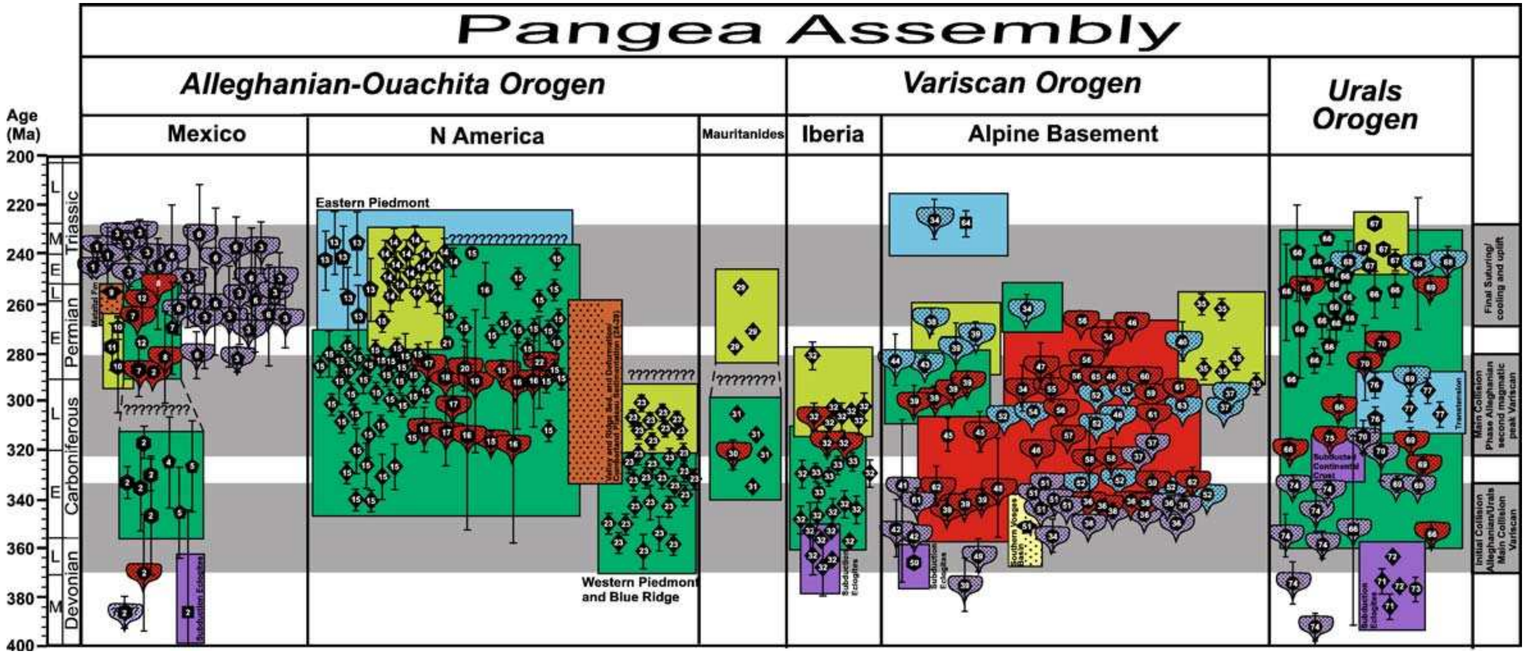

Fig. 8. Compilation of tectonostratigraphic data associated with tectonic events during the amalgamation of Pangea. The data indicate a correlation of events across the various orogens with initial collision of cratonic blocks occurring at around 360 Ma, followed by a thermal peak during the main crustal thickening episode between 320–280 Ma, followed by final suturing/cooling and uplift between 260–230 Ma. Numbers on data points refer to the following sources: 1 (Damon, 1975), 2 (Yanez et al., 1991), 3 (Torres et al., 1999), 4 (Araujo-Gómez and Arenas-Partida, 1986), 5 (López-Infanzón, 1986), 6 (Jacobo, 1986), 7 (Feo-Codecido et al., 1984), 8 (Grajales, 1988), 9 (Weber, 1997), 10 (Dennison et al., 1969), 11 (Carpenter, 1997), 12 (Elías-Herrera and Ortega-Gutiérrez, 2002), 13 (Hurley et al., 1960), 14 (Dallmeyer, 1982), 15 (Dallmeyer et al., 1986), 16 (Snoke et al., 1980), 17 (Fullager and Butler, 1979), 18 (Kish and Fullager, 1978), 19 (Russel et al., 1985), 20 (Weidemeyer and Spruill, 1980), 21 (Kocis et al., 1978), 22 (Fullager and Kish, 1981), 23 (Dallmeyer, 1988), 24 (Ferm, 1974), 25 (Horne et al., 1974), 26 (Secor et al., 1986b), 27 (Davis and Ehlrich, 1974), 28 (Hatcher et al., 1989), 29 (Dallmeyer and Villeneuve, 1987), 30 (Bassot and Caen-Vachette, 1983), 31 (Dallmeyer and Lecorche, 1990), 32 (Rodriguez et al., 2003), 33 (Escuder Viruete et al., 1998), 34 (Vavra et al., 1996), 35 (Brandmayr et al., 1995), 36 (Hegner et al., 2001), 37 (Schaltegger and Corfu, 1995), 38 (Quadt et al., 1999), 39 (Eichorn et al., 2000), 40 (Bussy et al., 1996), 41(Vavra and Hanson, 1991), 42 (Peindl and Höck, 1993), 43 (Söllner et al., 1991a), 44 (Loth et al., 1997), 45 (Cliff, 1981), 46 (Bussy and Hernandez, 1997), 47 (Pin, 1986), 48 (Eichorn et al., 1995), 49 (Kirsch et al., 1988), 50 (Bosse et al., 2000), 51 (Schaltegger et al., 1996), 52 (Schaltegger and Corfu, 1992), 53 (Sergeev et al., 1995), 54 (Bussy and von Raumer, 1993), 55 (Köppel and Grünenfelder, 1978), 56 (Bussy and Cadoppi, 1996), 57 (Bussy and Raumer, 1996), 58 (Bertrand et al., 1998), 59 (Schaltegger, 1997), 60 (Lippolt and Hess, 1996), 61 (Capuzzo and Bussy, 1998), 62 (Hess et al., 1995), 63 (Friedl et al., 1996), 64 (Voshage et al., 1987), 65 (Böhm, 1996), 66 (Echtler et al., 1997), 67 (Hetzel and Glodny, 2002), 68 (Kramm et al., 1983), 69 (Montero et al., 2000), 70 (Bea et al., 1997), 71 (Matte et al., 1993), 72 (Lennykh et al., 1995), 73 (Beane et al., 1996), 74 (Friberg et al., 2000), 75 (Hetzel and Romer, 1999), 76 (Glasmacher et al., 1999), 77 (Eide et al., 1997).

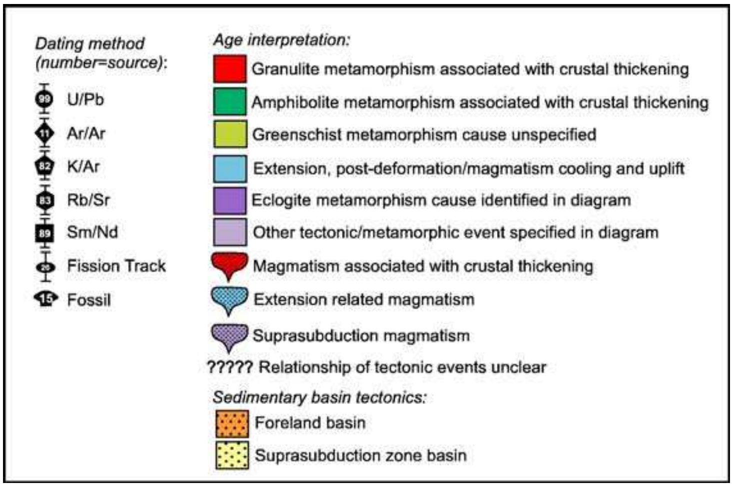

Fig. 8 (*continued*).

Tectonic and magmatic activity is also recorded in the Sierra de la Ventana fold belt in western Argentina (Rapela et al., 2003). Here a series of granites was intruded in large volumes beginning with the A-type Cerro Colorado granite at 531±4 Ma (U–Pb zircon) followed by the calc–alkaline San Mario Granite and Cerro del Corral rhyolite at 524±5 Ma (Rapela et al., 2003). Magmatic activity lasted for around 20 m.y. culminating with the intrusion of the La Ermita peralkaline rhyolite at 509±5 Ma (Fig. 6, U–Pb zircon; Rapela et al., 2003). These events have been correlated with the time equivalent Saldania Belt of South Africa where a similar series of A- to I-type granites of the Cape Granite Suite was intruded between *c.* 550 and 510 Ma (Armstrong et al., 1998; Rozendaal et al., 1999; Da Silva et al., 2000; Rapela et al., 2003). The tectonic history of both areas has been ascribed to a strike–slip dominated continental rift setting. However Rozendaal et al. (1999) and Rapela et al. (2003) noted that the I-type plutons of these suites resemble shoshonitic or high-potassium calc–alkaline volcanism typical of island arc or active continental margin settings suggesting that this rifting event may have occurred in a marginal basin setting above an active subduction zone.

4. Pangea events (400–200 Ma)

4.1. Pangea assembly

Pangea was the last supercontinent to have existed on Earth prior to the opening of the Atlantic, which heralded the development of the plate configuration seen today. Pangea formed by the amalgamation of Gondwana, Laurentia/Baltica and a continental mass comprising combined Siberia–Kazakhstan–Asia (Fig. 7). Amalgamation of these continental masses took place along: (i) the Alleghanian–Ouachita (Appalachian) Orogen in Mexico, North America and northwest Africa, suturing Laurentia and West Gondwana, (ii) the continuation of this orogen into the Variscan Orogen of Europe, suturing Europe and northwestern Gondwana, and (iii) the Urals Orogen, suturing the East European Craton and Siberia–Khazakhstan–Asia. The tectonic evolution of these orogens is complicated by later overprinting during the Alpine–Himalayan Orogeny in Europe and Asia and the opening of the Atlantic between South America and Africa. A full discussion of detailed events in the amalgamation of Pangea is available in Echtler et al. (1997), Li and Powell (2001), Stampfli and Borel (2002), and von Raumer et al. (2003). This section aims to summarise tectonostratigraphic data relating to temporal changes in the evolution of the Alleghanian–Ouchita, Variscan and Urals orogens.

4.1.1. Alleghanian–Ouachita Orogen

The Alleghanian–Ouchita Orogen stretches from the western margin of present day Mexico through Texas and along the eastern seaboard of the USA to Atlantic Canada.Its opposing margin is found in the Mauritanide–Basseride–Rokelide orogen and the Reguibat uplift of northwestern Africa (Figs. 7 and 9). The rocks within these regions record events related to late Palaeozoic collision of northwestern Gondwana with Laurentia.

In North America, Alleghanian–Ouchita events are most pronounced in the southeast section of the Appalachians in the Cumberland Plateau, Valley and Ridge and Piedmont provinces (Fig. 9; Hatcher et al.,

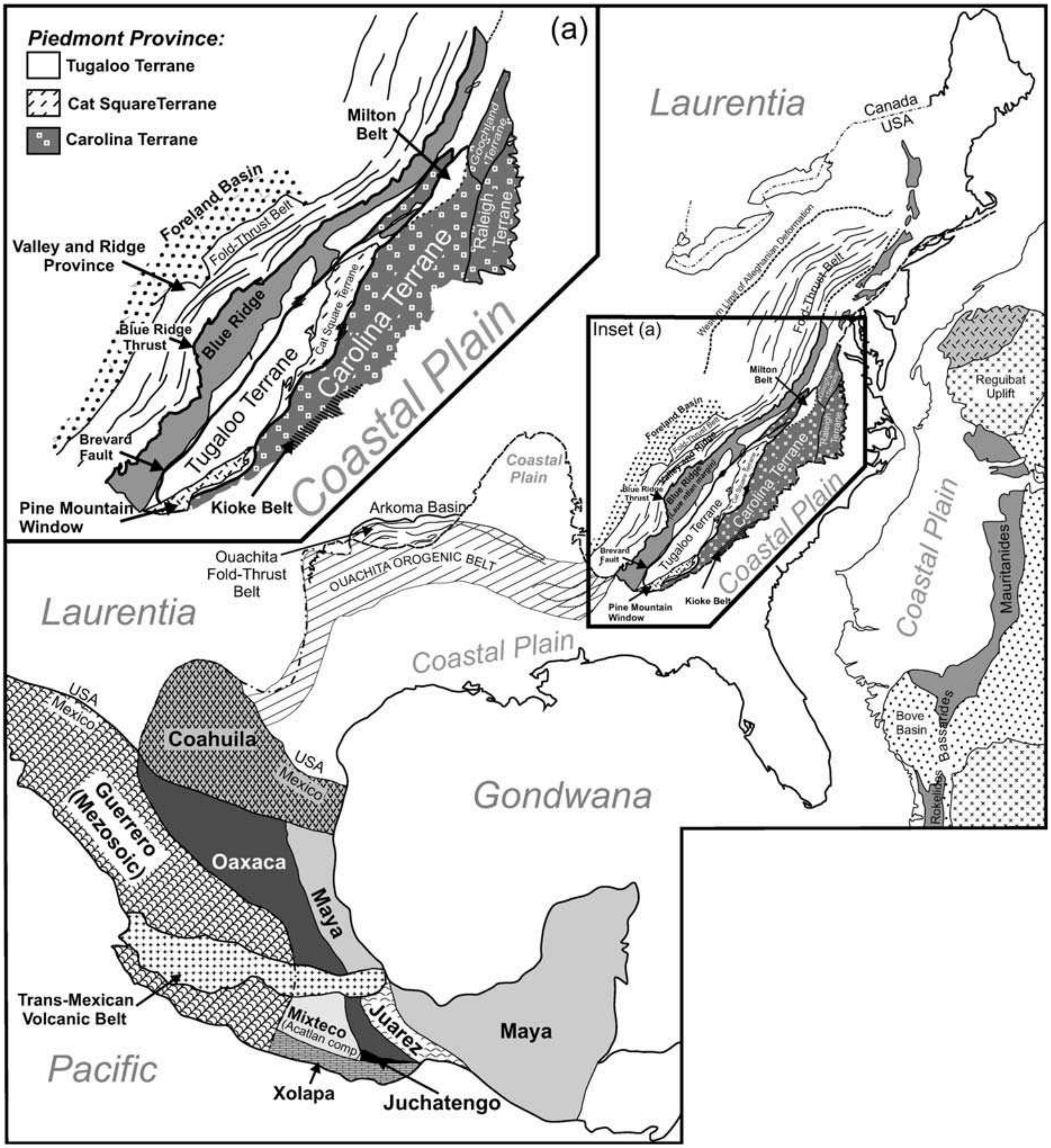

Fig. 9. Reconstruction of the Alleghanian–Ouachita Orogen showing tectonic terrane subdivisions in Mexico (after Dickinson and Lawton, 2001; Elías-Herrera and Ortega-Gutiérrez, 2002; after Yanez et al., 1991; Ortega-Gutierrez et al., 1995; Grajales-Nishimura et al., 1999), the United States of America (after Hatcher et al., 1989, 2004), and northwest Africa (after Dallmeyer and Lecorche, 1990; after Pique et al., 1987).

1989; Secor et al., 1986b), decreasing in intensity to the north into the Canadian Appalachians (van de Poll et al., 1995; Fallon et al., 2001; Gibling et al., 2002; Jutras et al., 2003). Events associated with the Alleghanian orogeny (*c.* 340–265 Ma) were first documented in the foreland basin in the Valley and Ridge Province and Cumberland Plateau of the southeast Appalachians (Fig. 8; Woodward, 1957), where a series of clastic wedges was deposited starting in the Late Mississippian (*c.* 320 Ma) and continuing to the youngest deformation affecting formations in the Dunkard Group in Ohio in the Late Pennsylvanian (ca. 285 Ma, Fig. 9; Davis and Ehlrich, 1974; ca. 285 Ma, Fig. 9; Rodgers, 1970; Ferm, 1974; Horne et al., 1974; Ross, 1986; Secor et al., 1986b). The Alleghanian orogeny is epitomised by successions in the Carboniferous Pocahontas basin that record a classic progressive unroofing sequence of emerging source terranes in the Piedmont to the east during the Namurian and Westphalian (Davis and Ehlrich, 1974; Ferm, 1974). The earliest deformation events are recorded by amphibolite to greenschist facies metamorphism and isotopic resetting of originally high-grade eastern Blue Ridge and western Piedmont province gneisses at around 362 Ma (Dallmeyer, 1988), slightly earlier than events in the lower grade Carolina Slate Belt (Fig. 9) to the east, which began around 340 Ma and continued until around 240 Ma (Fig. 9; Dallmeyer, 1982; Fig. 9; Hurley et al., 1960; Farrar, 1985; Dallmeyer et al., 1986; Secor et al., 1986a). In the northern Appalachians, in the Narragansett Basin of Rhode Island (Fig. 9), Pennsylvanian age sediments record a localised Barrovian style metamorphic event between 260 and 240 Ma thought to relate to Alleghanian deformation (Dallmeyer, 1982). During the main phase of deformation and metamorphism, a series of calc–alkaline granites was intruded in an arcuate belt stretching from Maryland to Georgia (Sinha and Zietz, 1982) that is dated between *c.* 325 and 280 Ma and is variably deformed as a result of the ongoing deformation (Kish and Fullager, 1978; Fullager and Butler, 1979; Snoke et al., 1980; Weidemeyer and Spruill, 1980; Fullager and Kish, 1981; Russel et al., 1985; Dallmeyer et al., 1986; Wintsch et al., 2003).

In Atlantic Canada, the Alleghanian period is represented by deposition and deformation of sedimentary sequences within fault bound sub-basins of the Maritimes Basin complex (van de Poll et al., 1995). Deformation was less intense than that experienced in the southern United States region and was dominated by transpressional strike–slip deformation resulting in folding of Carboniferous strata during the Late Palaeozoic (Nance, 1987; Fallon et al., 2001; Gibling et al., 2002; Jutras et al., 2003).

Contemporaneous events are recorded in the Mauritanide Orogen of western Africa (Fig. 9), where Neoproterozoic basement gneisses were overprinted

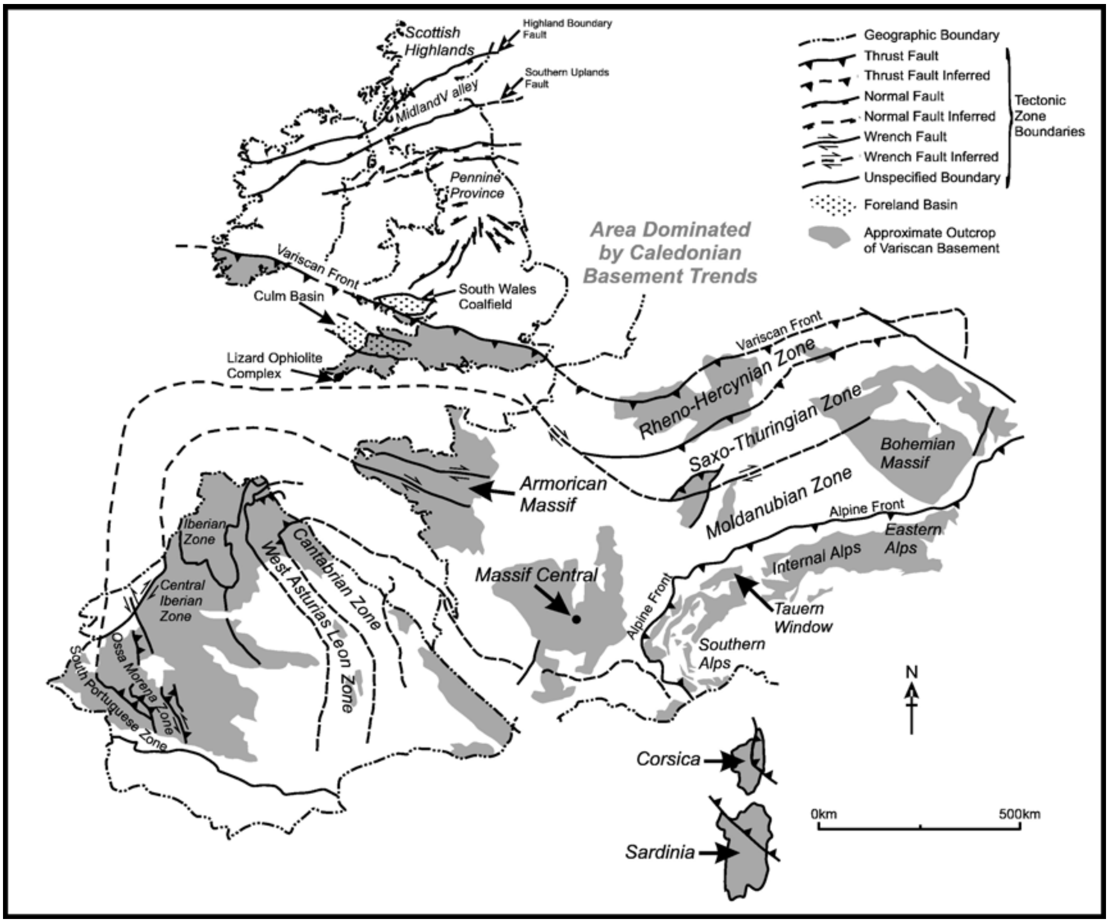

Fig. 10. Reconstruction of the European Variscan Orogen showing tectonic subdivisions (after Warr, 2000; von Raumer et al., 2003).

by amphibolite to greenschist facies fabrics between 300–280 Ma (Fig. 9; Dallmeyer and Lecorche, 1990; Fig. 9; Dallmeyer and Villeneuve, 1987). Drilling of subsurface gneisses in the Florida Peninsula and Gulf of Mexico has indicated that this area represents an extension of the Rokelides basement of Africa and records a similar Alleghanian deformation history to that of the Mauritanides exposures (Schlager et al., 1984; Dallmeyer et al., 1987; Dallmeyer and Villeneuve, 1987).

In the Mexican segment of the Ouachita belt (Fig. 9), the Laurentia–Gondwana collision is recorded by the amalgamation of several blocks interpreted as being sourced from both Gondwana and Laurentia prior to final collision of the two main continental bodies (Anderson and Schmidt, 1983; Ortega-Gutierrez et al., 1995; Carpenter, 1997; Ortega-Gutierrez et al., 1999; Dickinson and Lawton, 2001; Elías-Herrera and Ortega-Gutiérrez, 2002; Harry and Londono, 2004). The associated deformation and magmatic events are most clearly defined on the margins of the Proterozoic Oaxaca and Palaeozoic Acatlán complexes (Fig. 8 and 9; Yanez et al., 1991; Elías-Herrera and Ortega-Gutiérrez, 2002). The Acatlán complex consists of multiply deformed metasedimentary rocks, granites and eclogites that are often compared to units in the Appalachians (Yanez et al., 1991). Although these metasedimentary sequences have been tied to the Iapetus Ocean (Ortega-Gutierrez et al., 1999) recent work suggests that they accumulated along the Gondwana margin of the Rheic Ocean (Murphy et al., 2006). They are correlated with unmetamorphosed strata of Ordovician age which together contain detrital zircons with ages indicating derivation from the Oaxacan Complex and Amazon craton (Murphy et al., 2006). The Esperanza granites, which intrude both the Acatlán and Oaxaca terranes are of Ordovician age (480–440 Ma) and along with coeval mafic bodies (e.g. Middleton et al., 2006) may be part of a bimodal assemblage related to opening of the Rheic Ocean. Late Devonian to Carboniferous (360–330 Ma) deformation and metamorphism (López-Infanzón, 1986; Yanez et al., 1991; Carpenter, 1997; Elías-Herrera and Ortega-Gutiérrez, 2002; Middleton et al., 2006) is related to initial collision between Gondwana and Laurentia during the amalgamation of Pangea. A suite of plutons in the Acatlán complex, the Toltopec Granites, containing zircons with concordant ages of 287±2 Ma is consistent with ages from elsewhere in Mexico suggesting widespread granite magmatism at this time (Feo-Codecido et al., 1984; Araujo-Gómez and Arenas-Partida, 1986; Grajales, 1988; Yanez et al., 1991; Dickinson and Lawton, 2001). The boundary between the Oaxaca and Acatlán complexes is known as the Caltepec Fault zone, here high-temperature metamorphism and deformation led to the generation of migmatites at 275±1 Ma, interpreted as dating the peak of metamorphism, and intrusion of the Cozahuico Granite at *c.* 260 Ma (Fig. 8, U–Pb zircon; Carpenter, 1997; Fig. 8, U–Pb zircon; Elías-Herrera and Ortega-Gutiérrez, 2002). Further evidence of a metamorphic peak at around 275 Ma is provided by Rb–Sr studies of muscovite schists in the Sierra del Carmen metamorphic suite (Fig. 8, 277±10 Ma Muscovite–Feldspar–Apatite–Whole Rock isochron; Carpenter, 1997). At the same time a complex transformation was taking place with the establishment of an active margin on the eastern edge of the newly amalgamated blocks resulting in widespread calc–alkaline magmatism between *c.* 280–210 Ma in a narrow belt from the Juchatengo suite in the southeast to the Coahuila region stitching many of the terrane boundaries (Jacobo, 1986; Torres et al., 1999). These events suggest prolonged deformation in this region, which is interpreted to represent the final collisional stages between Gondwana and Laurentia (Yanez et al., 1991; Dickinson and Lawton, 2001; Elías-Herrera and Ortega-Gutiérrez, 2002).

4.1.2. Variscan Orogen

The Variscan Orogen stretches across the whole expanse of Europe covering virtually the same area as the modern Alpine Orogen, with the inclusion of areas in the Iberian region of Spain and also southern England and northern France (Figs. 7 and 10). The record of amalgamation of peri-Gondwanan terranes and a combined Laurentia/Baltic craton within the Variscan basement rocks is complicated by younger overprinting but several recent studies have managed to see through the Alpine events and reconstruct a history for this time (e.g. Schaltegger and Corfu, 1995; Schaltegger, 1997; Stampfli and Borel, 2002; e.g. von Raumer, 1998; von Raumer et al., 2002, 2003).

The Iberian Massif of Spain records several metamorphic and magmatic events during Variscan orogenesis. In the northwest Iberian Massif, the Malpica–Tui complex preserves evidence of four different deformational events defined by $^{40}Ar/^{39}Ar$ studies of white micas from eclogite to amphibolite facies rocks (Rodriguez et al., 2003). The earliest deformation records high-pressure conditions in phengite white micas from eclogite-facies pelitic gneisses with ages in the range *c.* 370–365 Ma (Fig. 8), considered to represent the onset of continental margin subduction (Rodriguez et al., 2003). Dates from gneisses interpreted to represent the contact between the subducting plate and overriding plate give ages in the range *c.* 350–340 Ma (Fig. 8), interpreted as the time at which the

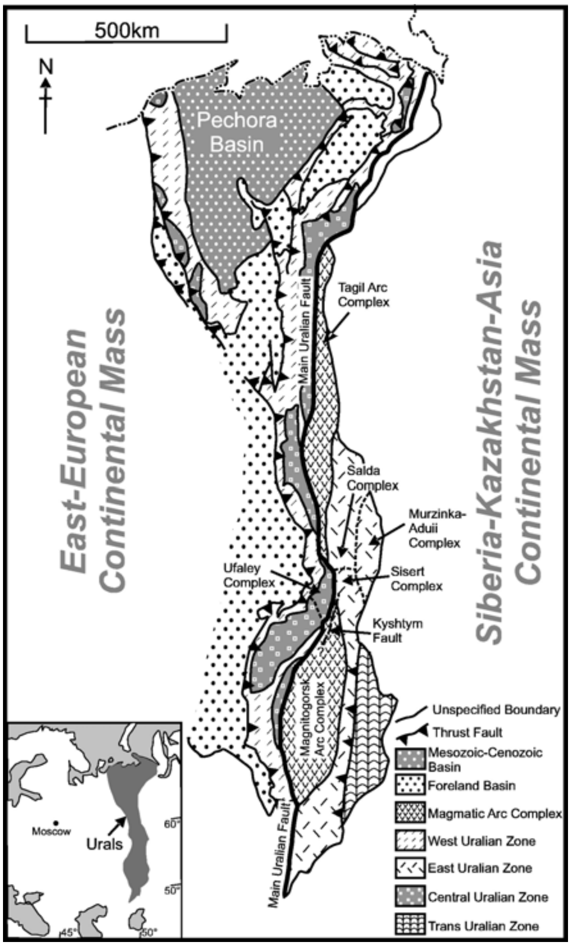

Fig. 11. Map of the Urals Orogen showing tectonic subdivisions (Echtler et al., 1997; Hetzel et al., 1998; Hetzel and Romer, 1999; Friberg et al., 2000; Hetzel and Glodny, 2002).

two plates became attached (Rodriguez et al., 2003). This peak of metamorphism was rapidly followed by exhumation of the subducted margin resulting in retrograde amphibolite metamorphism between *c.* 325 and 315 Ma, during which time convergence between the two plates continued (Fig. 8 Rodriguez et al., 2003). A final period of convergence and granite intrusion is recorded between *c.* 310 and 280 Ma producing alkaline orthogneisses and syntectonic granites after which deformation in the area ceased (Rodriguez et al., 2003). The central Iberian Massif records a very similar history in the Sierra de Guadarrama, where the Berzosa–Riaza shear zone underwent syncollisional extension with low-*P*/high-*T* metamorphism resulting in monazite growth between *c.* 337 and 326 Ma (Escuder Viruete et al., 1998).

The basement of the modern Alpine Orogen (Fig. 10) records a complex history of Variscan deformation, metamorphism and magmatic events. The Tauern window of Austria, records three distinct pulses of magmatism associated with the Variscan orogeny. The final stages of subduction related magmatism prior to collision of the European Craton and Avalonia is recorded by Late Devonian juvenile calc–alkaline precursors of the Zwölferkogel granodioritic gneiss, which has yielded a U–Pb zircon age of 374±10 Ma (Fig. 8; Eichorn et al., 2000). Similar rocks have been identified within the Armorican terrane (Fig. 10) with an age of 362±7 Ma (Kirsch et al., 1988) and within the Schwartzburg anticline and Central Bohemian Batholith of the Moldanubian zone (Fig. 10; Eichorn et al., 2000; Fig. 10; Kosler et al., 1993; Janousek et al., 1995). These granitoids are succeeded by collision-related crustal melt granites of the Hochweissenfeld gneiss at 342±5 Ma, syncollisional extensional deformation in the Falkenbachlappen gneiss at 343±6 Ma and by the Augengneiss of the Felbertauern complex at 340±4 Ma (Eichorn et al., 2000). Each of these intrusions is interpreted to be related to the collision of the peri-Gondwanan Tauern terrane with Laurentia/Baltica (von Raumer, 1998; Eichorn et al., 2000; von Raumer et al., 2002, 2003). Similar events are reported in the external Alpine basement of Switzerland where subaerial clastic sediments and pyroclastic tuffs were deposited in syncollisional basins between *c.* 345 and 335 Ma (Schaltegger and Corfu, 1995) accompanied by intrusion of high-K magmas and peak metamorphism at *c.* 337–333 Ma within basement gneisses (von Raumer, 1998). This was followed by a period of uplift, cooling and erosion of the gneisses before eruption of pyroclastic protoliths of the Schönbachwald, Hauschartenkpopf and Peitingalm gneisses between 300 and 279 Ma (Eichorn et al., 2000), which were laid down unconformably on a 334 Ma equivalent of the Augengneiss of Felbertausern (Vavra and Hanson, 1991). Intrusion of calc–alkaline I-type granites accompanied this basin formation and volcanism between *c.* 299 and 295 Ma (Eichorn et al., 2000). The calc–alkaline nature of these rocks has led to some confusion regarding the lack of a subducting margin at this time, but Schaltegger and Corfu (1995) and Schaltegger (1997) argue that similar magmatism in the external massifs was the result of post-orogenic uplift and adiabatic decompression of previously subducted oceanic lithosphere underplated into the lower continental crust. Variscan compression had ceased in the Tauern window and external massifs by *c.* 270 Ma with the emplacement of S- and A-type post-orogenic granites interpreted to have been intruded during a period of extension in continental wrench zones (Schaltegger,

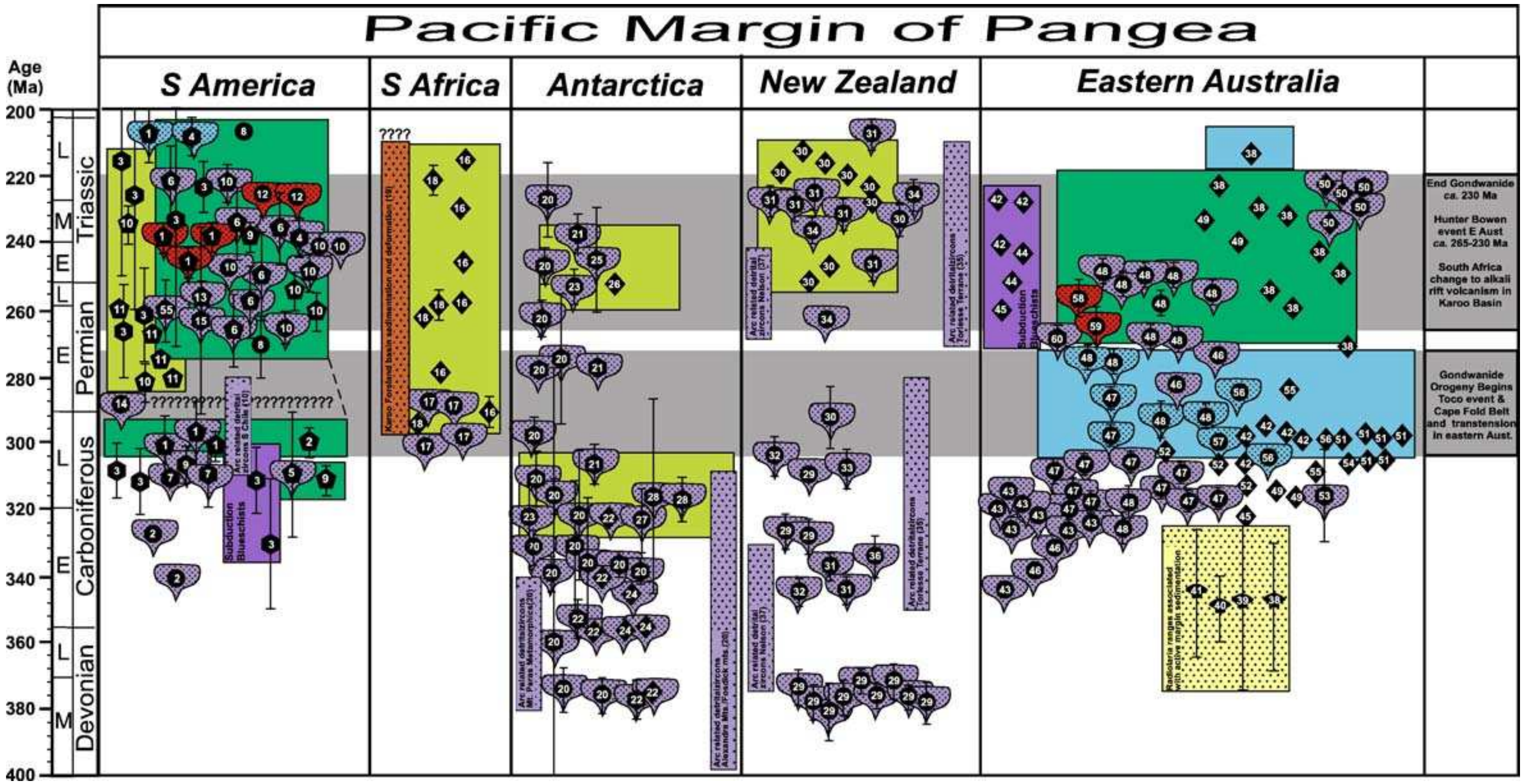

Fig. 12. Compilation of tectonostratigraphic data associated with tectonic events along the Pacific margin of Pangea between 400–200 Ma. The data indicate a temporal correlation of events along the margin as margin extension in eastern Australia is synchronous with the inset of the Gondwanide orogeny along the rest of the margin around 305–300 Ma. A switch from margin extension to compression took place in Eastern Australia at *c.* 265 Ma ceased at 230 Ma synchronous with the end of compression along the rest of the margin. Numbers on data points refer to the following sources: 1 (Nasi et al., 1985), (Höckenreiner et al., 2003), 3 (Hervé, 1988), 4 (Rapela and Kay, 1988), 5 (Ribba et al., 1988), 6 (Linares et al., 1980), 7 (Rex, 1987), 8 (Ramos, 2000), 9 (Hervé et al., 1985), 10 (Martin et al., 1999), 11 (Varela et al., 1985), 12 (Noble et al., 1997), 13 (Llambias et al., 2003), 14 (Damm et al., 1990), 15 (Brook et al., 1987), 16 (Hälbich et al., 1983), 17 (Bangert et al., 1999), 18 (Gresse et al., 1992), 19 (Cole, 1992), 20 (Pankhurst et al., 1998b), 21 (Pankhurst et al., 1996), 22 (Adams, 1987), 23 (Pallais et al., 1993), 24 (Richard et al., 1994), 25 (Adams et al., 1995), 26 (Mukasa, 1995), 27 (Mukasa et al., 1994), 28 (Stump, 1995), 29 (Muir et al., 1996), 30 (Mortimer et al., 1999), 31 (Kimbrough et al., 1994), 32 (Muir et al., 1994), 33 (Bradshaw et al., 1996), 34 (Tulloch et al., 1999), 35 (Adams et al., 1998), 36 (Gibson and Ireland, 1996), 37 (Adams et al., 2002), 38 (Holcombe et al., 1997a), 39 (Aitchison, 1990), 40 (Ishiga, 1990), 41 (Spiller, 1993), 42 (Little et al., 1993), 43 (Roberts et al., 2003), 44 (Ashley and Brownlow, 1993), 45 (Fukui et al., 1993), 46 (Kimbrough et al., 1993), 47 (Roberts et al., 2004), 48 (Roberts et al., 1996), 49 (Sliwa et al., 1993), 50 (Stephens et al., 1993), 51 (Little et al., 1992), 52 (Watanabe et al., 1990), 53 (Webb and McDougall, 1968), 54 (Little et al., 1995), 55 (Jones et al., 1996), 56 (Allen et al., 1994), 57 (Dirks et al., 1993), 58 (Gulson et al., 1990), 59 (Collins et al., 1993), 60 (Flood et al., 1988).

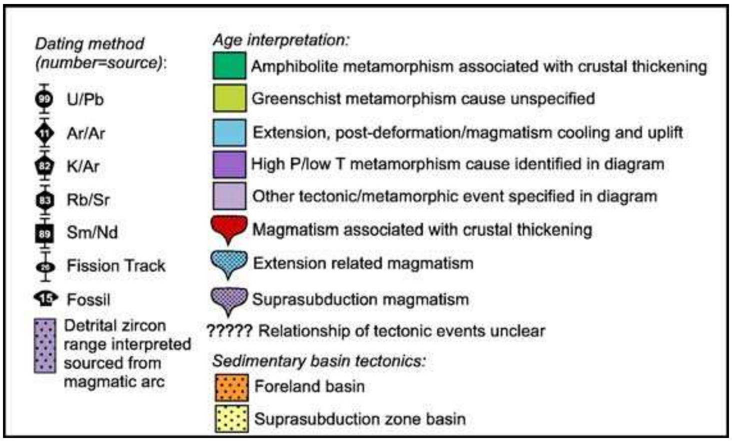

Fig. 12 (*continued*).

1997) or during the initial stages of the Alpine orogeny (Quadt et al., 1999).

A similar multi-phase history is recorded in the Southern Alps of Italy (Fig. 10), where gneisses in the Ivrea Zone record granulite metamorphism at 355 Ma followed by decompression melting in pelitic and psammitic layers between *c.* 296 and 260 Ma (Fig. 11; Vavra et al., 1996). The final metamorphic event recorded in these rocks occurred at 226±5 Ma (Vavra et al., 1996). This is significantly later than those in the Tauern window and external massifs, suggesting some diachroneity in deformation within the Variscan orogen.

The Internal Alps massifs (Fig. 10) record final subduction related magmatism at *c.* 350 Ma followed by collision-related metamorphism and granite emplacement in the Southern Vosges and Schwartzwald between *c.* 345 and 330 Ma (Fig. 11; Hegner et al., 2001; Fig. 11; Schaltegger et al., 1996). Syncollision extensional basins are also found in the Southern Vosges region between *c.* 340 and 330 Ma (Schaltegger et al., 1996) suggesting that these were a common feature along the length of the Variscan Orogen. Late-Carboniferous to Permian (*c.* 320–290) extension and granite emplacement is recorded in the Mont-Mort area and represents the final stages of Variscan collision-related deformation in the Internal Alps (Bussy and Cadoppi, 1996; Bussy et al., 1996; Bertrand et al., 1998; von Raumer, 1998).

4.1.3. Urals Orogen

The Urals Orogen was created during the amalgamation of eastern Europe with a combined Siberia–Kazakhstan–Asia continent during the formation of Pangea (Figs. 7 and 11). As with the Variscan Orogen, the rocks of the Urals record a multi-phase deformational, metamorphic and magmatic history. However, the history of the Urals is more clearly preserved due to a lack of subsequent overprinting.

Supra-subduction magmatic arc volcanism is recorded in the Magnitogorsk Arc (Fig. 11) of the Urals through the Devonian until *c.* 350 Ma (Fig. 8; Echtler et al., 1997). Towards the end of this period, at *c.* 370–355 Ma eclogites and blueschists were formed along the Main Uralian fault (Figs. 8 and 11), which marks the main suture between the largely magmatic arc terranes of the Siberian–Kazakhstan–Asia continent and the continental margin of the East European craton (Hetzel et al., 1998). Immediately following this high-pressure metamorphism a period of granulite facies high-temperature metamorphism

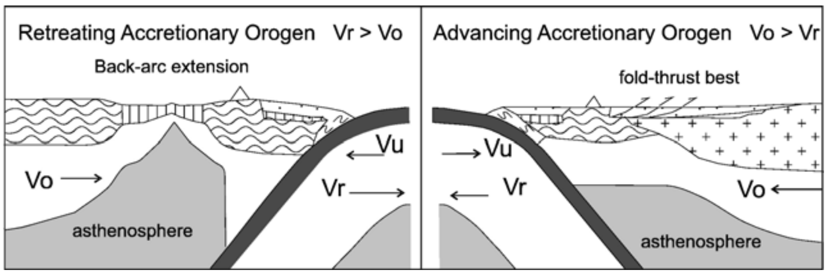

Fig. 13. Summary of tectonic events associated with amalgamation of Gondwana and coeval margin activity.

and granite intrusion is recorded in the Salda metamorphic complex at *c.* 350–335 Ma (Figs. 8 and 11) that is interpreted to represent the onset of collision of Eastern Europe and the Kazakhstan–Siberia–Asia continent (Friberg et al., 2000). A second phase of convergent deformation immediately followed by extensional exhumation is recorded in the Ufaley complex in the Middle Urals, part of the East European continental margin. This complex was subducted to eclogite facies conditions and then rapidly exhumed through amphibolite facies by 316±1 Ma (Fig. 8) when a granite was intruded across the foliation in the Ufaley gneiss (Fig. 11 Hetzel and Romer, 1999). The period of extensional exhumation was accompanied by the intrusion of several tonalites, trondhjemites and granodiorites between *c.* 320 and 315 Ma (Bea et al., 1997). There then followed a period of relative tectonic and magmatic quiescence, after which renewed compression resulted in a magmatic pulse creating several granites throughout the Urals between 290 and 240 Ma (Bea et al., 1997; Echtler et al., 1997). The final stages of deformation in the Urals were influenced by a change in relative compression direction resulting in orogen parallel strike–slip fault development and high-temperature metamorphism and mylonite generation in the Kyshtym Fault between *c.* 240 and 229 Ma (Fig. 8), after which deformation along the Main Uralian Fault ceased (Hetzel and Glodny, 2002).

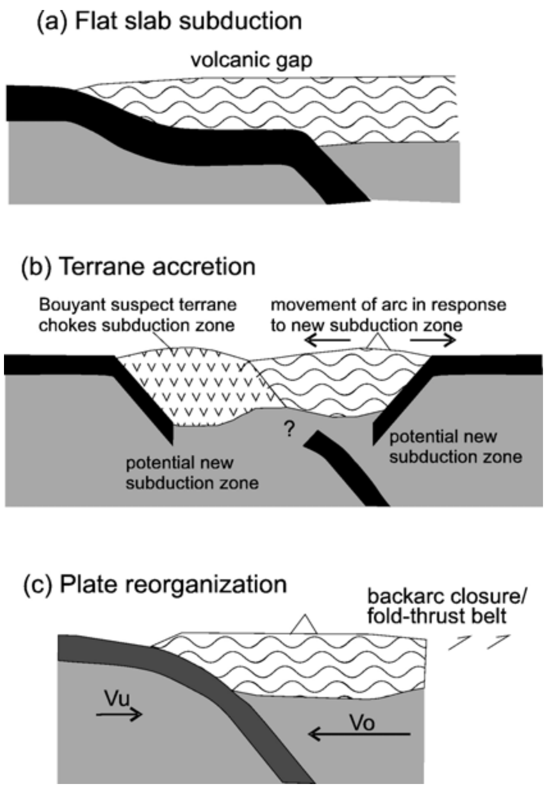

Fig. 14. Summary of tectonic events associated with amalgamation of Pangea and coeval margin activity.

4.2. Pacific margin of Pangea

The Pacific margin of Gondwana remained a coherent margin throughout the formation of Pangea following the late Neoproterozoic to early Palaeozoic amalgamation of Gondwana. Events along the margin during this period are punctuated by the Gondwanide orogeny that affected the whole of the Terra Australis Orogen (Figs. 2 and 12).

The East Gondwana margin remained an active subduction–accretion margin throughout the Palaeozoic (Cawood and Leitch, 1985; Scheibner, 1998; Foster et al., 2005a) and the first events associated with the Gondwanide orogeny involved a complex interplay of compression and transtension between about 305 Ma and 230 Ma (Leitch, 1988; Veevers, 2004). Tectonic events associated with the Gondwanide orogeny are recorded in accretionary prism rocks of the Tablelands Complex in the New England region of eastern Australia, where they are marked by mid-crustal deformation and metamorphism along with emplacement of S-type granites (e.g. Hillgrove Suite) at around 305 Ma (Collins et al., 1993). Deformation has been related to contraction in the New South Wales segment (Dirks et al., 1993) and extension in the Queensland segment (Little et al., 1995). A phase of extension (probably sinistral transtension) occurred between 300 and 270 Ma, resulting in the generation of the Sydney–Bowen and Barnard basins (Cawood, 1982; Leitch, 1988; Holcombe et al., 1997b; Veevers, 2004). The main phase of deformation, referred to locally as the Hunter–Bowen Orogeny (Carey and Browne, 1938), occurred between 265 and 230 Ma and is well developed throughout the New England region (Leitch, 1969; Collins, 1991; Veevers, 2004). Deformation extended west (decreasing in intensity and age) into the Sydney–Bowen basin, which evolved into a foreland system (Fergusson, 1991; Korsch, 2004) with the oldest detritus shed from the uplifting welt of the New England region dated at about 270 Ma (Hamilton, 1986; Roberts et al., 1996). At the same time, major oroclinal bending occurred in the east (Cawood, 1982; Korsch and Harrington, 1987; Murray et al., 1987; Goss and Cawood, 2005). In New Zealand, the Hunter–Bowen event is poorly documented, but Mortimer et al. (1999) have proposed that it was related to accretion of the Brook Street Terrane. Gondwanide deformation of variable intensity and distribution is recognized throughout

West Antarctica and the adjoining Cape Fold Belt of southern Africa on the basis of stratigraphic and geochronological data (Storey et al., 1987; Trouw and De Wit, 1999; Johnston, 2000). In the Ellsworth–Whitmore Mountains, Permo-Triassic Gondwanide deformation resulted in upright to inclined folds with axial planar cleavage that are inferred to have formed in a dextral transpressive environment (Curtis, 1998).

Following Ordovician (?) margin extension, the West Gondwana margin records passive margin sedimentation until the Mid-Carboniferous when it developed into a continental arc setting above an active subduction zone, as indicated by intrusion of batholiths in the Chilean Frontal Cordillera at *c.* 320–310 Ma (Mpodozis and Kay, 1992). This margin switch immediately predates the onset of the Gondwanide Orogeny which includes the Late Carboniferous Toco event and the mid-Permian San Rafael (Sanrafaelic) event (Bahlburg and Hervé, 1997; Ramos, 2000). The Toco event involved folding and melange disruption of turbidite strata as young as Late Carboniferous to Early Permian with an upper age limit provided by the emplacement of *c.* 310–290 Ma plutons into the folded turbidites (Bahlburg and Hervé, 1997). The San Raphael event is marked by intense folding and thrusting, resulting in a pronounced angular unconformity between Late Carboniferous to Early Permian turbidites and the extensive Permo-Triassic Choiyoi Volcanics (Kay et al., 1989; Ramos, 2000; Ramos and Aleman, 2000). The Ventana Fold Belt, inboard of the Andes, is characterized by NNE verging fold and thrust belt, the development of which is contemporaneous with that of the Sauce Grande foreland basin (Trouw and De Wit, 1999). Deformation occurred between about 280 and 260 Ma on the basis of K–Ar ages and is inferred to have taken place in a dextral transpressional environment (Cobbold et al., 1991). Ramos (2000) proposed that the accretion of Patagonia drove deformation in the Ventana and Cape fold belts (cf. Pankhurst et al., 2006).

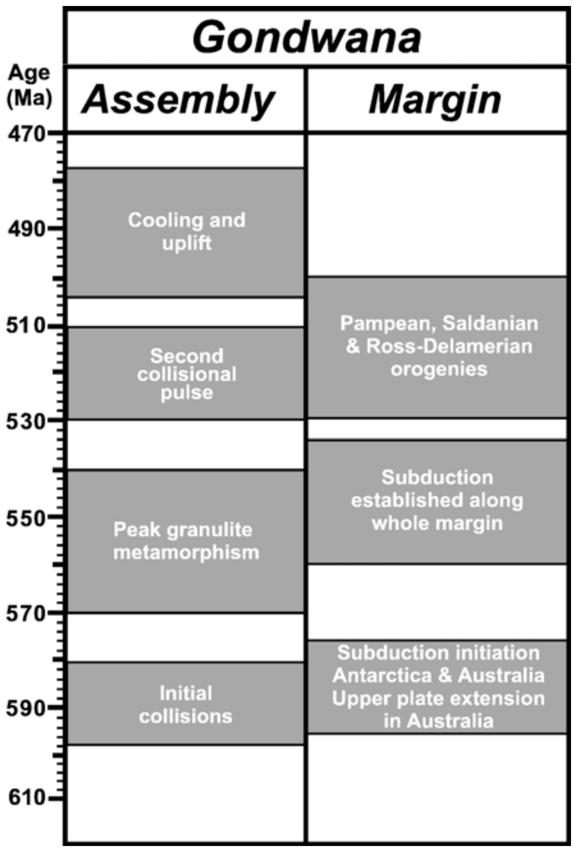

Fig. 15. Accretionary orogen types: for the retreating-type the velocity of slab retreat (Vr) for the underriding plate (Vu) is greater than that of the overriding plate (Vo), whereas for the advancing orogen the velocity of the overriding plate is greater.

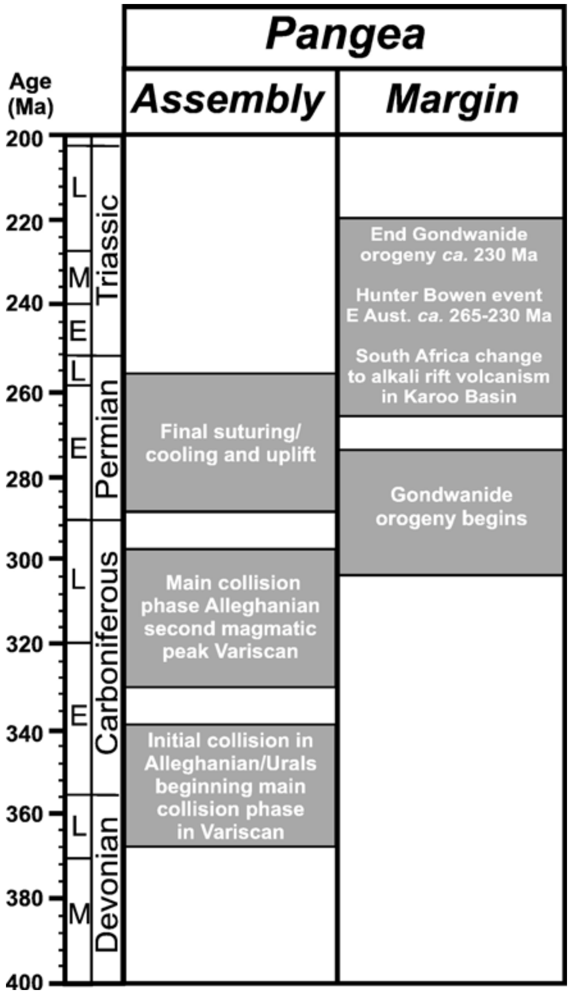

Fig. 16. Possible modes of plate coupling leading to orogenesis in accretionary orogens.

Permo-Triassic orogenesis along the Pacific margin was accompanied by the widespread emplacement of I-type granites marking a new outboard, convergent margin accretionary cycle. The main products of this margin are exposed in the old fore-arc and subduction complex of the New England Fold Belt (Cawood, 1984), along the Median Tectonic Zone in New Zealand (Mortimer et al., 1999), and in Marie Byrd Land (Mukasa and Dalziel, 2000), the Antarctic Peninsula (Vaughan and Storey, 2000) and South America (Ramos and Aleman, 2000).

Permo-Triassic deformation events also affected the southern margin of South Africa where basement rocks were uplifted to form the Cape Fold Belt and associated Karoo Foreland Basin. Ash fall tuffs in the Dwyka Formation deposited between *c.* 300 and 288 Ma are interpreted to provide evidence of arc related magmatism along the margin of Gondwana at this time (Bangert et al., 1999), but the distance of the Cape Fold Belt and Karoo basins from the margin is under debate (Cole, 1992; Gresse et al., 1992, Thomas, 1993 #294; Visser, 1992). Compressional activity within the Cape Fold Belt itself is dated by propagation of deformation into the Karoo basin, resulting in a number of basin inversion events at *c.* 300, 260, 250 and 220 Ma and greenschist grade foliation growth in pelitic sediments (Hälbich et al., 1983; Gresse et al., 1992). Gresse et al. (1992) suggested that these events can be correlated with unconformities within the Karoo sedimentary sequence, but Frimmel et al. (2001) argue that these unconformities are related to older pre-Cape events and thus cannot be correlated. Nonetheless, the greenschist cleavage forming events are not in dispute and clearly identify deformation in the Cape Fold belt between 300 and 220 Ma, synchronous with events along the rest of the Gondwana Pacific margin.

5. Discussion

5.1. Driving compression in accretionary margins

Accretionary orogens can be broken into two end-member types (Fig. 13; cf. Uyeda, 1982), *retreating* and *advancing*. Retreating orogens undergo long-term extension in response to lower plate retreat resulting in forearc accretion and opening of back arc basins (e.g. West Pacific). Advancing orogens develop in an environment in which the overriding plate is advancing

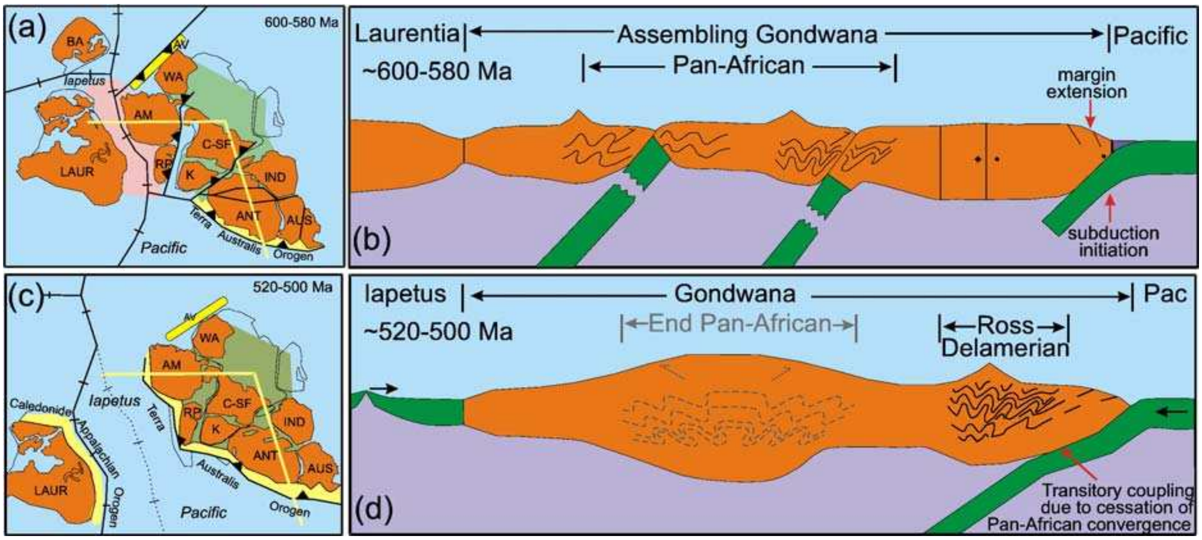

Fig. 17. Schematic model illustrating the mechanism to induce transient coupling at the plate margin by transfer of shortening from the Pan-African orogens in the latter stages of amalgamation of Gondwana. (a) Tectonic plate configuration *c.* 600–580 Ma, bold line indicates line of section shown in (b). (b) Schematic cross section illustrating subduction initiation at the Pacific Margin of Gondwana caused by plate reconfiguration after initial collision of continental blocks along Pan-African orogens. Rapid roll-back at the newly formed subduction zone produces a retreating-mode accretionary orogen and results in margin extension in Australia. (c) Tectonic plate configuration *c.* 520–500 Ma, bold line indicates line of section shown in (d). (d) Schematic cross section illustrating onset of compression at Pacific margin resulting in the Ross–Delamerian, Pampean and Saldanian Orogenies. After cessation of activity in the Pan-African orogens, shortening was transferred to the subducting margin causing a change to advancing-mode and thus transient coupling across the plate boundary and compression of the upper plate. In (a) and (c) orange is stabilized continental crust, pale green is Pan-African orogens, pink is rifted continental crust related to Iapetus opening, yellow is Terra Australis and Caledonide–Appalachian orogens and dark yellow is area of Avalon (Av) and related terranes. Yellow line is position of schematic cross sections (b) and (d). In (b) and (d), orange is continental crust and green is oceanic crust.

towards the downgoing plate, as exemplified by the modern westward motion of the North and South American plates advancing over the eastward moving Pacific plate (Russo and Silver, 1996). This has resulted in the accretion (and strike–slip motion) of arc and microcontinental ribbons previously rifted off the upper plate and extensive retro-arc fold and thrust belts. The mode of plate convergence may switch between that of an advancing or retreating orogen such as has been documented in Eastern Australia (Holcombe et al., 1997b; Collins, 2002; Jenkins et al., 2002) and the Permo-Triassic belts of South America (Ramos and Aleman, 2000). The process of *tectonic switching* of kinematic reference frames appears to be intimately related to accretionary orogenesis (Collins, 2002), but the driving force for switching from an extensional to compressional regime is poorly understood.

5.2. Detecting convergent margin coupling and orogenic driving mechanisms

Possible mechanisms for coupling within accretionary orogens include; subduction of buoyant oceanic lithosphere (*flat-slab subduction*); accretion of buoyant lithosphere (*terrane accretion*); and *tectonic plate reorganization* (Fig. 14). These mechanisms punctuate a regime of ongoing plate convergence, resulting in transitory coupling across the boundary and orogenesis. Each mechanism has the following predictable consequences for the geological record.

Flat-slab subduction reflects the effect of subduction of young (e.g. oceanic ridge) and/or buoyant (e.g. oceanic plateau) lithosphere (e.g. Murphy et al., 2003; e.g. Murphy et al., 1998). Effects should be spatially limited to the region of the flat-slab/accretion zone (Fig. 14a), which should result in short-lived orogenesis and/or diachronous events that migrate along the convergent margin in harmony with the subducted plate movement vector (Saleeby, 2003). Flat-slab subduction in modern settings (e.g. Andes, Ramos et al., 2002) also produces characteristic geochemical signatures (Kay and Mpodozis, 2002) in associated igneous rocks including adakitic compositions (Samaniego et al., 2002). In addition igneous activity should cease in the previously established arc and may migrate inboard. If impinging oceanic plateaus were the cause of increased coupling, then there should be evidence of accreted parts of the buoyant plateau preserved in the accretionary complex (e.g. Cawood, 1990). Mid-Ocean Ridge subduction is a specific case of flat-slab subduction which may involve a major change in plate convergence vectors between the upper plate and two different subducting plates, with the change in plate convergence vectors separated by a period of heating and igneous intrusion in the forearc or in the trench. The result is deformation that is diachronous, and will include structural, magmatic, thermal, and metallogenic fingerprints (Bradley et al., 2003).

Terrane accretion (Fig. 14b) should also be spatially limited to the region of terrane collision and, similar to flat-slab subduction, should have short-lived/diachronous effects. For terrane accretion to be a driving mechanism of orogenesis the accreted terrane must be brought to the upper plate margin on the subducting lower plate and therefore must be exotic to the plate margin. If the terrane is (para)autochthonous (e.g. rifted micro-continental ribbon or arc) then its accretion is a consequence of orogenesis rather than a driving force (e.g. Vaughan and Livermore, 2005). This can be tested using provenance and or palaeomagnetic studies to detect if the accreted terrane is truly exotic (see summary in; Vaughan et al., 2005). During terrane accretion magmatic arc activity may migrate first inboard as terrane accretion causes contraction and then jump outboard as a new subduction zone is established. Although the North American Cordillera is often cited as the type example of terrane related orogenesis, recent work has shown that most, if not all, suspect terrains are upper plate fragments of American affinity (e.g. English and Johnston, 2005; e.g. Monger and Knokleberg, 1996; Johnston, 2001), and are superficial with no deep crustal roots (Snyder et al., 2002).

Tectonic plate reorganization (Fig. 14c) involving changes in convergence direction, including rapid increases in the absolute motion of the overriding plate, will affect the length of the orogen/plate boundary synchronously and reflect widespread and possibly long-term changes in orogenic character. Plate reorganization may be inter-orogen in extent. For example, major plate reorganization associated with increased spreading rate in the Pacific during the mid-Cretaceous (Sutherland and Hollis, 2001) resulted in increased plate buoyancy and major submarine flood magmatism, and is interpreted to have resulted in pan-Pacific margin tectonic and metamorphic effects (Vaughan, 1995; Vaughan and Livermore, 2005). This process contrasts markedly with those detailed above and is thus readily distinguishable by correlating timing and kinematics of events along the margin and also with coeval events around the world that may have led to plate reorganization.

5.3. Amalgamating supercontinents and marginal compression

The data presented for assembly of Gondwana and Pangea indicate that the timing of collisional orogenesis between amalgamating continental bodies was synchronous with subduction initiation and contractional orogenesis within accretionary orogens located along the margins of these supercontinents. In the case of Gondwana, final assembly occurred between *c.*570 and 510 Ma (Fig. 15), amalgamating the various components of East and West Gondwana with final cooling and uplift complete by 490 Ma (Fig. 3). This was coeval with a switch from passive margin sedimentation to convergent margin activity along the Pacific margin of the supercontinent (Fig. 15). Timing of subduction initiation along the Pacific margin ranges from 580–550 Ma evidenced by the first appearance of arc derived detrital zircons in the upper Byrd group sediments and the oldest supra-subduction zone plutons along the Antarctic segment of the margin (Fig. 6). A phase of extension marked by supra-subduction zone ophiolite generation at 535–520 Ma is preserved in greenstone successions in eastern Australia and immediately precedes Ross–Delamerian contractional orogenesis between 520 and 490 Ma (Fig. 6) inboard of the plate margin, which coincides with the cessation of collisional orogenesis between the amalgamating blocks of Gondwana (Fig. 15). Supra-subduction zone igneous activity was continuous throughout this period indicating that subduction was ongoing along the Gondwana margin (Fig. 6).

The final stages of assembly of the Pangean supercontinent occurred between *c.*320 and 250 Ma (Fig. 8). Major plate boundary reorganization during this time (Valencio et al., 1983; Kay et al., 1989) was accompanied by regional orogenesis along the Pacific margin. The East Gondwanan margin segment experienced extension and strike–slip activity from *c.*305 Ma until *c.*270 Ma, after which convergence along the plate margin was re-established, marked in eastern Australia by the migration of arc magmatism in response to a migration of the plate margin (Cawood, 1984; Jenkins et al., 2002). Synchronous with this phase of plate re-adjustment was the Gondwanide Orogeny (305–230 Ma), which affected the entire Pacific margin of Pangea (Figs. 2 and 12).

These temporal relations between interior collisional orogeny, associated with assembly of continental blocks, and marginal accretionary orogens during supercontinent amalgamation (Figs. 15 and 16), suggest a linked history between interior and exterior processes, which we believe is related to global plate kinematic adjustments (Fig. 14c) rather than localised mechanisms such as terrane accretion or flat-slab subduction (Fig. 14a, b). Orogenesis in accretionary orogens occurs in the absence of colliding bodies during ongoing subduction and plate convergence and must therefore be driven by a transitory coupling across the plate boundary. Correspondence of coupling along the margin of a supercontinent with assembly of the supercontinent and termination of convergence in its interior suggests that it may reflect plate re-adjustments involving a temporary phase of increased relative convergence across the plate boundary.

Fig. 17 shows a schematic model illustrating how this may be explained in terms of the amalgamation of Gondwana, but is conceptually equally valid at the time of Pangea assembly. Prior to collision of continental blocks in Gondwana, oceans existed that were freely subducting allowing shortening to be taken up along the subducting margins and a passive margin to exist along the Pacific margin of Gondwana (Fig. 17a). When initial collision of continental blocks occurred in the Pan-African orogens, the effect would be to rapidly reduce the convergence velocity between these bodies (Fig. 17b) and, if a model of a constant radius Earth is assumed, potentially cause the plate configuration to change between Gondwana and the Pacific Ocean lithosphere, thereby initiating subduction and thus transferring convergence to the plate margin (Fig. 17b). The balance between motion across the Pan-African orogens and the newly initiated subduction zone would stabilize convergence and plate movement for the period of Pan-African orogenesis. However, when convergence ceased along the Pan-African orogens at *c.* 520–500 Ma as Gondwana was finally amalgamated (Fig. 3), plate motions would again be altered as convergence could no longer be accommodated in this region (Fig. 17d). The result would be to transfer convergence to the free moving subducting margins of Gondwana, such as the Pacific margin, thereby causing an increase in convergence velocity and coupling between the overriding and subducting plates, and producing the Ross–Delamerian, Saldanian and Pampean orogenesis in the upper plate (Fig. 17d). It is expected that the coupling effects would be transitory as the margin adapted to the change in convergence rate and would return to a steady subduction configuration resulting in a decrease of coupling across the boundary.

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