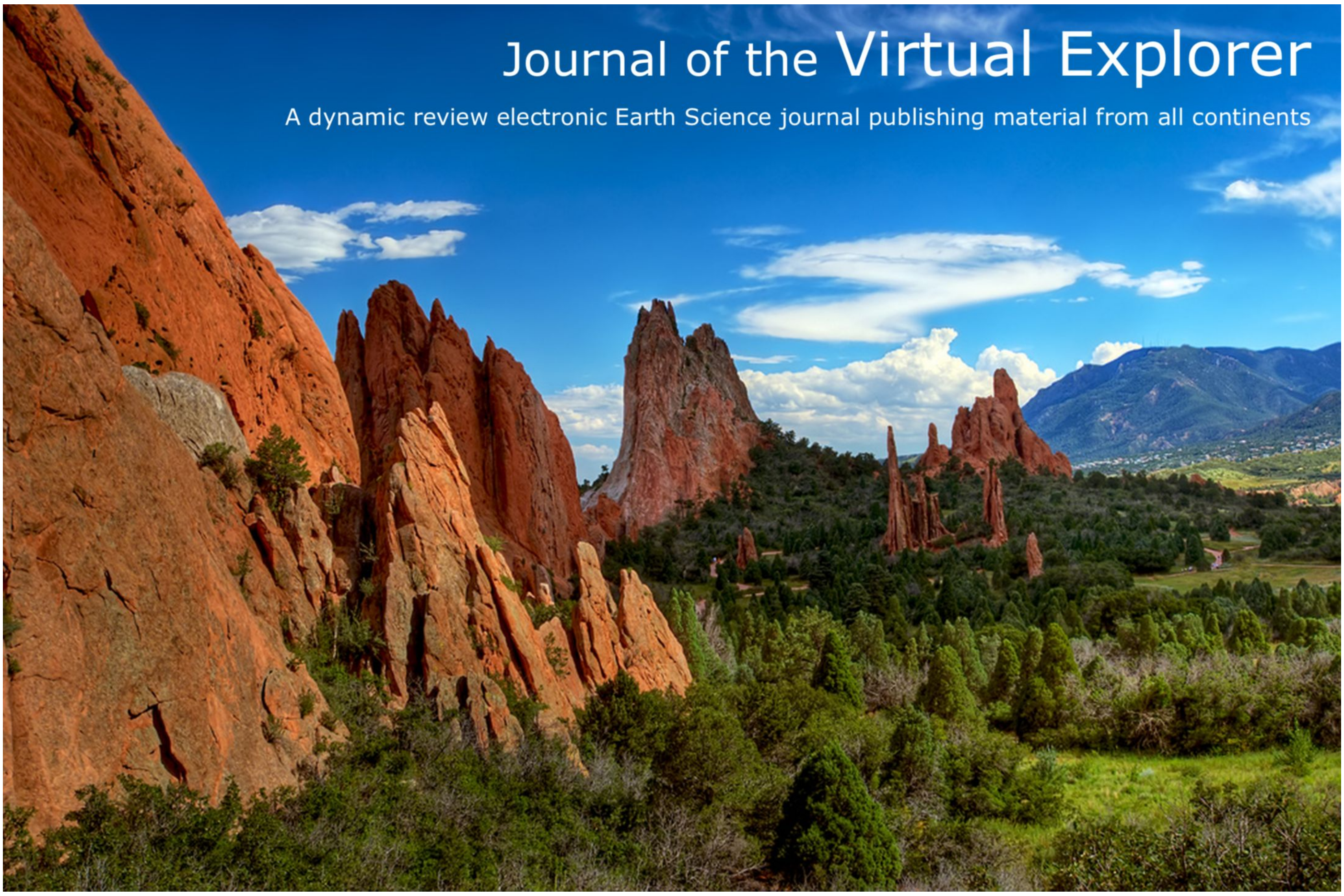

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Oroclinal Bending in the southern New England Orogen (eastern Australia): a geological field excursion from Brisbane to Sydney

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Abstract: A series of tight bends are recognised in the southern New England Orogen in the area between Brisbane and Newcastle. This field excursion will provide an opportunity to become familiar with this spectacularly contorted orogen, to evaluate alternative geometrical interpretations, and to discuss possible tectonic models. The geometry of the New England oroclines is controversial, with alternative models suggesting a structure of 2, 3 or 4 oroclines. The tectonic setting responsible for these curvatures is largely unknown. During the four-day field excursion, we will explore the different tectonic elements that outline the oroclinal structure, including rocks that originated in a Devonian-Carboniferous convergent margin (subduction complex, forearc basin and volcanic arc) and a folded belt of early Palaeozoic serpentinites and high-pressure rocks. An emphasis will be given to exposures of early Permian rocks, predominantly S-type granitoids and rift-related sedimentary rocks, which most likely originated in a backarc setting during the early stages of oroclinal bending.

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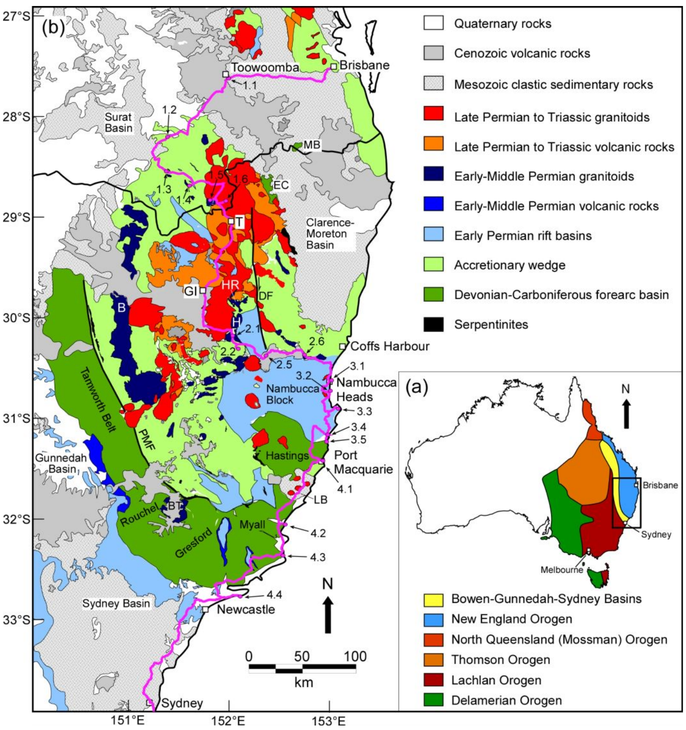
Introduction

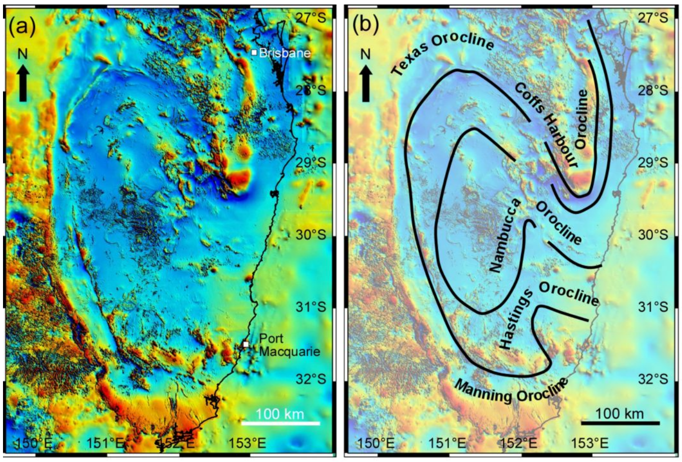
The Palaeozoic orogenic system in eastern Australia, in the area between Brisbane and Sydney, forms a number of tight bends (oroclines) (Figs. 1 and 2). The origin, and even the exact geometry, of these oroclines are still debated. Some authors argue that the structure is limited to a single Z-shaped bend known as the Texas-Coffs Harbour Orocline (Murray *et al.*, 1987; Lennox and Flood, 1997; Offler and Foster, 2008). Others have suggested a more complex orogenic structure, comprising three bends (Korsch and Harrington, 1987) or even four bends (Cawood *et al.*, 2011b; Glen and Roberts, 2012; Rosenbaum, 2012; Rosenbaum *et al.*, 2012). The existence of such tight curvatures raises many questions related to the origin of the oroclines, their 3D geometry and continuation at depth, and the mechanism of oroclinal bending. These questions, which are largely still unresolved, have major implications on the way we interpret the evolution of the Australian continent and the way we understand deformation of the continental margin.

Figure 1. Location maps.



(a) Location map showing the Australian continent, the Tasmanides (coloured areas, after Glen, 2005) and the New England Orogen; (b) Geology of the southern New England Orogen. The field itinerary and stops are indicated in the purple line and circles, respectively (see also Figure 3). B, Bundarra Granite; BT, Barrington Tops Granodiorite; DF, Demon Fault; EC, Emu Creek Block; H, Hillgrove plutonic suite; LB, Lorne Basin; GI, Glen Innes; HR, Henry River Granite; MB, Mt Barney Inlier; PMF, Peel-Manning Fault System; T, Tenterfield.

Figure 2. Magnetic maps.



(a) Total magnetic intensity image of the southern New England Orogen (after Milligan *et al.*, 2010). (b) Schematic interpretation of the major orogenic curvatures in the southern New England Orogen.

In this fieldtrip, we will examine the different geological entities that define the New England oroclines. We will see the pre-oroclinal tectonic elements, which predominantly originated in a Devonian to Carboniferous supra-subduction environment. We will also examine an earlier (Ordovician?) suture, which appears in a folded serpentinite belt. We will look at the assemblage of, possibly backarc-related, early Permian rocks, which seem to provide crucial information on the timing of oroclinal bending and associated geodynamics. We will make observations on a later phase of Permo-Triassic deformation and magmatism, and will discuss the origin of younger sedimentary and volcanic rocks.

The oroclinal structure, particularly of the Texas-Coffs Harbour Orocline, is clearly recognised in aeromagnetic images (Figure 2) and is expressed in the curvature of bedding and structural fabrics (Korsch, 1981; Lennox and Flood, 1997; Aubourg *et al.*, 2004; Li *et al.*, 2012a). The opposite-vergent oroclinal pair farther south, the Manning and Nambucca oroclines (Figure 2b), are somewhat more elusive. Their existence is supported by a number of independent criteria, such as the map-view curvature of early Permian granitoids, the contorted serpentinite belt and the arrangement of pre-oroclinal forearc basin terranes (Glen and Roberts, 2012; Rosenbaum, 2012; Rosenbaum *et al.*, 2012). A number of authors, however, are sceptical, arguing that further evidence is required in order to support the existence of the Manning and Nambucca oroclines (Offler and Foster, 2008; Lennox *et al.*, 2012). This fieldtrip itinerary includes a number of key observations related to this debate.

The fieldtrip is planned for 4 days. In order to allow an orogen-scale discussion, the itinerary covers a relatively large distance (total of >1400 km, including detours) and involves a few long driving legs (Figure 3). The first day is dedicated predominantly to the Texas Orocline. In the second day, we will visit early Permian, possibly backarc-related, S-type granitoids, and will traverse the high-grade Wongwibinda metamorphic complex. In the third day, we will see multiple generations of deformation in the early Permian Nambucca Block. Lastly, in the fourth day, we will examine serpentinites and high-pressure rocks at Port Macquarie, and will then travel south towards Sydney through a number of outcrops in the eastern limb of the Manning Orocline.

Figure 3. Field Trip itinerary.

Map showing the fieldtrip itinerary (source: Google Map). The itinerary is also available in *Google Earth (kmz)* format.

Geological Setting

The geology of eastern Australia is considered to be a type-example of a subduction-related accretionary orogen (Cawood and Buchan, 2007). The resulting assembly of Palaeozoic orogens, normally referred as the Tasmanides (Glen, 2005), makes a large component of the Australian continent (Figure 1a) and generally becomes younger from west to east. The New England Orogen is the easternmost and youngest component of the Tasmanides. It has developed in a supra-subduction environment from Devonian to Triassic, although recycled components from an earlier orogen (Cambrian and Ordovician) are also found (Aitchison *et al.*, 1994; Fukui *et al.*, 1995).

The southern New England Orogen occupies the area between Brisbane and Newcastle (Figure 1b). The majority of the rocks in this area are associated with the forearc of an arc related to a Devonian-Carboniferous subduction zone (Leitch, 1974). These rocks are overlain by early Permian sedimentary basins, which were deposited simultaneously with the emplacement of S-type granitoids (Shaw and Flood, 1981; Korsch *et al.*, 2009). A later stage of voluminous calc-alkaline, I-type, magmatism took place from late Permian to Triassic (260-220 Ma), possibly in a continental arc setting (see Bryant *et al.*, 1997).

The Devonian to Carboniferous forearc region is expressed in forearc basin rocks (predominantly of the Tamworth Belt) and the subduction complex rocks of the Tablelands Complex (Figure 1b). The Tamworth Belt comprises fluvial to shallow marine sedimentary rocks deposited on a shelf that was deepening from west to east (Roberts and Engel, 1987). The southern continuation of the Tamworth Belt is the Rouchel, Gresford and Myall blocks. Other blocks attributed to the forearc basin are the Hastings Block (Korsch, 1977; Roberts *et al.*, 1995b) the Emu Creek Block (Cross *et al.*, 1987), and possibly also the small inlier of Mt Barney in southeast Queensland (Olgers *et al.*, 1974) (Figure 1b). The Tablelands Complex consists of variably deformed and metamorphosed deep marine volcaniclastic turbidites, cherts and argillites, mafic volcanic rocks, and olistostromal deposits (Leitch and Cawood, 1980; Cawood, 1982; Fergusson, 1984). Metamorphic conditions vary from prehnite-pumpellyite/lower greenschist-facies to high-grade (amphibolite-facies) metamorphic complexes (Korsch, 1978; Stephenson and Hensel, 1982; Phillips *et al.*, 2008; Danis *et al.*, 2010; Craven *et al.*, 2012).

The other elements of the Devonian to Carboniferous convergent margin, namely the arc and back-arc regions, are not exposed in the southern New England Orogen, with the exception of a few occurrences of arc-related rocks in external parts of the Tamworth Belt (e.g. Nerong Volcanics). The majority of the volcanic arc was originally positioned west of the Tamworth Belt (Korsch, 1984), but was subsequently overthrust by the forearc basin (Glen and Roberts, 2012).

The contact between the Tamworth Belt and the Tablelands Complex is a tectonic contact, the Peel-Manning Fault System (Figure 1b), along which, serpentinites and high-pressure rocks are exposed (Benson, 1913; Aitchison *et al.*, 1994; Och *et al.*, 2003). These rocks, which are Cambrian-Ordovician in age, are the oldest component in the New England Orogen, with metamorphic ages that are roughly similar to rocks from the Lachlan Orogen in Victoria (Phillips and Offler, 2011, and references therein).

Early Permian rocks in the southern New England Orogen are predominantly S-type granitoids and clastic rift-related sedimentary successions. The former occurs mainly in two granitic suites, the elongate Bundarra Granite in the west, and the granitoids of the Hillgrove plutonic suite (Figure 1b). The two granitic suites, along with a number of additional granitoids, were emplaced simultaneously at 295-290 Ma (Cawood *et al.*, 2011a; Rosenbaum *et al.*, 2012), and their map-view structure delineates the shape of the New England oroclines (Figure 1b). Granite emplacement was contemporaneous with the early development of back-arc extensional basins (Korsch *et al.*, 2009), possibly in response to subduction rollback (Jenkins *et al.*, 2002; Rosenbaum *et al.*, 2012). Clastic sedimentation occurred in a number of peripheral basins (Sydney, Gunnedah and Bowen Basins), which were subsequently inverted into foreland basins, as well as in other more internal basins (e.g. Nambucca Block, Figure 1b). Although these basins are thought to represent an early Permian backarc environment (Korsch *et al.*, 2009), the exact architecture of the convergent margin remains speculative due to the lack of information on the early Permian arc and forearc regions.

Following the emplacement of the early Permian S-type granitoids, magmatic activity ceased in the southern New England Orogen from 280 to 260 Ma (with the exception of the ~267 Ma Barrington Tops Granodiorite, Figure 1b). Magmatism resumed at ~260 Ma, forming the broad NE-SW New England batholith of subduction-related granitoids and volcanic rocks (1b). At about the same time (from ~265 Ma to ~235 Ma), the region was subjected to ~E-W contractional deformation, normally referred as Hunter-Bowen Orogeny. At the end of this period of contractional deformation, from 235 Ma onwards, the focus of magmatic activity shifted eastward, possibly in response to asymmetric subduction rollback (Li *et al.*, 2012b).

During the Mesozoic, eastern Australia experienced widespread basin formation and sedimentation. Jurassic and Cretaceous continental clastic sediments and coal measures were deposited in the Clarence-Moreton and

Surat Basins unconformably on top of the earlier rocks (Figure 1b), thus obscuring the connection between the southern and the northern segments of the New England Orogen.

The youngest geological components in the New England region are Cenozoic basalts (Figure 1b), which are predominantly lava flows derived from a number of central volcanoes. The origin of this volcanism has been attributed to a hotspot beneath the northward moving Australian plate (Vasconcelos *et al.*, 2008).

Day 1

The excursion begins with a trip from Brisbane westwards to the city of Toowoomba (~700 m a.s.l). Brisbane itself is built on Palaeozoic low-grade metamorphic rocks that originated in a Devonian-Carboniferous subduction complex (equivalent to the Tablelands Complex in the southern New England Orogen). These rocks are locally overlain by Triassic and Jurassic rocks (Brisbane Tuff and Ipswich Basin). Travelling east, we encounter Mesozoic sedimentary rocks of the Clarence-Moreton Basin, before climbing onto a high plateau made of Cenozoic basalts.

Stop 1.1. Toowoomba

GPS coordinates (lat/long): -27.578959°/151.987349°

Distance from Brisbane: 125 km

The first stop is located at the eastern edge of Toowoomba and provides an excellent view to the east. The steep topography, which marks the edge of the Diving Range, is the expression of the passive margin escarpment that has retreated due to erosion. The actual escarpment in this locality is made of Cenozoic basalts that belong to the western flank of a shield volcano (Cohen, 2012). Beneath the basalts is the Mesozoic Clarence-Moreton Basin. These rocks are characterised by flat-lying clastic sedimentary successions, mostly Jurassic sandstones, siltstone, mudstones, shales and conglomerates, and some prominent coal measures. The basin started to develop in the latest Triassic and continued to subside and to accumulate sediments during the Mesozoic.

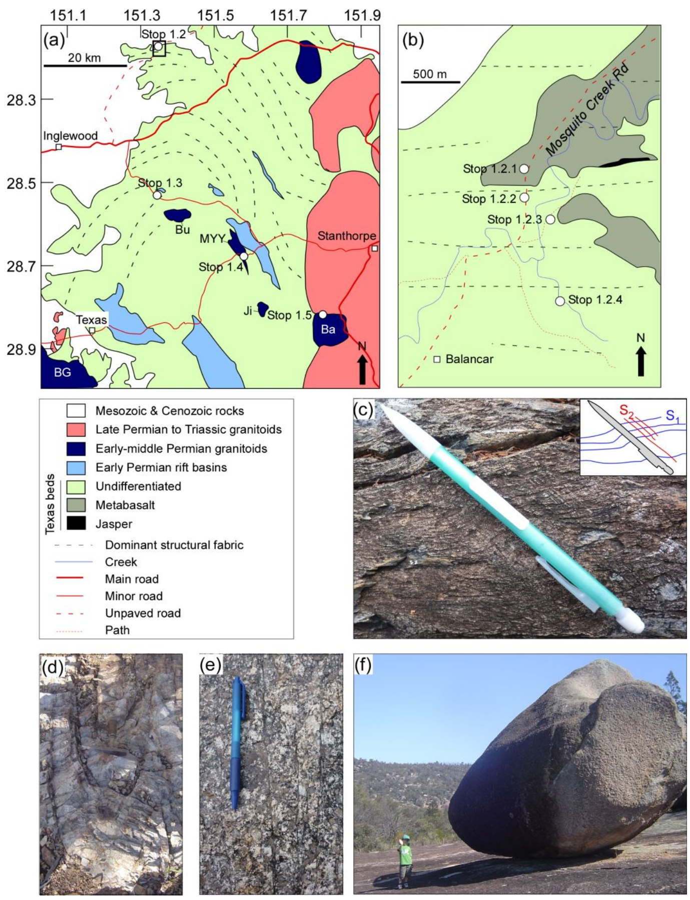
Stop 1.2. The Texas beds in Mosquito Creek

Distance from previous stop: 114 km

Travelling south from Toowoomba, we approach exposures of the Palaeozoic New England Orogen. The rocks in this area are called the Texas beds (Figure 4a) and are part of the Devonian-Carboniferous subduction complex (Tablelands Complex). Their lithology is dominated by volcaniclastic turbidites (lithic arenite and mudstone) with minor chert, limestone and mafic volcanic rocks (Donchak *et al.*, 2007). The source of the well-bedded arenites is typically associated with silicic volcaniclastic detritus. Rocks were metamorphosed in lower greenschist facies conditions and record penetrative ductile deformation.

The area around Mosquito Creek (Figure 4b) is located in the hinge of the Texas Orocline, which is one of the most noticeable oroclinal features in eastern Australia (Figure 1c). The orocline is clearly defined by variations in the orientation of bedding and structural fabrics within the Texas beds, changing from NW-striking in the eastern limb, to NE-striking in the western limb (Figure 4a). In this locality, the dominant bedding and structural fabric is E-W (Figure 4b), as expected in the hinge zone.

Figure 4. Geology of the Texas Orocline.



(a) Geological map of the Texas Orocline and locations of Stops 1.2 to 1.5 (modified after Li *et al.*, 2012a). Ba, Ballandean Granite; BG, Bundarra Granite; Bu, Bullaganang Granite; Ji, Jibbinbar Granite; MYY, Mt You You Granite. Box indicates location of the inset map in Mosquito Creek. (b) Simplified geological map of the hinge of the Texas Orocline in

Mosquito Creek (modified after Li and Rosenbaum, 2010) and locations of the fieldtrip sites. (c) Low strain crenulation cleavage in the Texas beds (Stop 1.2.3). (d) Isoclinal F1 fold in the Texas beds (Stop 1.3). (e) Spaced NW-SE cleavage in the early Permian Ballandean Granite (Stop 1.5). (f) A large boulder of the Stanthorpe Granite in Girraween National Park (Stop 1.6). The scale is a (very charming) 5-year-old child.

Stop 1.2.1. Metabasalt

GPS coordinates: -28.187486°/151.346113°

The rock in this locality is a greenschist facies metabasalt comprising chlorite, plagioclase and calcite. It is locally associated with pillow basalts, indicating submarine extrusive mafic volcanism, and commonly appears in association with thin-bedded layers of jasper and quartzite (Donchak *et al.*, 2007). Spaced E-W sub-vertical cleavage can be recognised in this outcrop, parallel to the dominant structural fabric in the hinge zone of the orocline. Note that the map-scale orientations of the metabasalt bodies (NE-SW and SE-NW; Figure 4b) are oblique to the local ~E-W bedding orientation.

Stop 1.2.2. Slaty cleavage in the Texas beds

GPS coordinates: -28.189587°/151.346121°

Distance from previous stop: 0.23 km

Strong slaty cleavage is recognised in the meta-sedimentary quartz-rich beds. Rocks incorporate abundant micro-scale quartz veins, which are isoclinally folded (F1) with axial planes parallel to the dominant slaty cleavage (S1). The dominant structural fabric (S1) is consistently sub-vertical E-W.

Stop 1.2.3. Low strain crenulation cleavage

GPS coordinates: -28.191222°/151.348358°

Distance from previous stop: 0.78 km

Please note that this outcrop, as well as the next outcrop (Stop 1.2.4), are located in a private property. Entrance requires permission from the owners.

The rocks in this outcrop show overprinting relationships between the dominant S1 slaty cleavage and a later structural fabric. The latter appears as a weak crenulation cleavage (Figure 4c). Whether or not this structural fabric is directly related to the formation of the Texas Orocline is unclear. In any case, the lack of a more penetrative structural fabric parallel to the axial plane of the orocline (NNW-SSE) is somewhat surprising and may indicate that oroclinal bending involved little strain (Li *et al.*, 2012a).

Stop 1.2.4. Pre-oroclinal quartz veins and vergence relationship

GPS coordinates: -28.197473°/151.349296°

Distance from previous stop: 0.87 km

This relatively extensive creek bed outcrop shows overprinting relationships between early N-S-trending quartz veins and the dominant E-W structural fabric (correlative to the dominant S1 slaty cleavage in Stop 1.2.2). The S1 fabric is parallel to the axial plane of minor folds and cuts the quartz veins, indicating that these veins formed prior to the development of the pre-oroclinal F1 folds. A very low angle is recognised between the orientations of the sub-vertical bedding (S0, steeply dipping to the south) and the dominant fabric (S1, steeply dipping to the north), allowing the determination of vergence relationship associated with the pre-oroclinal F1 folds.

Stop 1.3. Texas beds in Waroo

GPS coordinates: -28.535430°/151.350809°

Distance from previous stop: 59 km

Travelling south from Mosquito Creek (4a), we see occasional outcrops of the relatively monotonous succession of the Texas beds. Minor folds are not common. However, in this locality we can see meso-scale F1 isoclinal folds (Figure 4d). The NE-SW strike of the axial plane is parallel to the orientation of the western limb of the Texas Orocline. Overprinting F2 folds are not recognised.

Stop 1.4. Mt You You Granite

GPS coordinates: -28.675064°/151.581383°

Distance from previous stop: 38.5 km

The elongated NNW-SSE Mt You You Granite is located in the eastern limb of the Texas Orocline (Figure 4a) and crosses the Stanthorpe-Texas Rd in Pike Creek. Rocks are exposed in the creek bed, so accessibility is sometimes limited due to flooding.

The composition of this pluton is rather heterogeneous, varying from granite to biotite monzogranite, syenogranite and minor hornblende-biotite monzogranite. In this locality, the granite includes mafic and intermediate enclaves. A recent U-Pb SHRIMP zircon age from this locality yielded an age of 294.0±2.9 (Rosenbaum *et al.*, 2012). This age is similar to the age of the other small early Permian plutons in the eastern limb of the Texas Orocline (Bullaganang, Mt You You, Ballandean and Jibbinbar) and is only slightly older than the ~290 Ma Bundarra Granite, which is located farther to the southwest (Figure 4a). It seems, therefore, that the Mt You You Granite is part of a belt of early Permian granitoids that is folded around the Texas Orocline.

Stop 1.5. Ballandean

GPS coordinates: -28.813810°/151.794491°

Distance from previous stop: 64.5 km

A short detour from the New England Highway will allow us to look at the Ballandean Granite (Figure 4a), which is another S-type early Permian (~295 Ma) pluton on the eastern limb of the Texas Orocline. The granite is slightly deformed and is characterised by a spaced cleavage oriented NW-SE (Figure 4e).

Visiting this area gives us an opportunity to explore some of the local wineries. This region in southeast Queensland, known as the Granite Belt, has more than 50 high-altitude (700-1200 m a.s.l) vineyards. Main wine production varieties include Chardonnay, Verdelho, Merlot, Cabernet Sauvignon and Shiraz.

Stop 1.6. Girraween

GPS coordinates: -28.833448°/ 151.935713°

Distance from previous stop: 21 km

Girraween National Park is situated in the Granite Belt immediately north of the Queensland-NSW border (Figure 3). The rocks are primarily Lower-Middle Triassic I-type granites. The largest of which is the Stanthorpe Granite, which is beautifully exposed in Girraween National Park in the form of large boulders, inselbergs and tors (Figure 4f). If time permits, it is worthwhile taking one of the short walks starting from Bald Rock Creek camping area, including the Junction (5 km return, GPS -28.830167°/151.929215°) and the Pyramid (3.4 km return, GPS -28.821884°/151.944438°).

The Stanthorpe Granite is a composite granitic suite, which together with the Ruby Creek Granite, makes the northern part of the New England Batholith (Shaw and Flood, 1981). Its composition is predominantly leucogranitic with >74% SiO_2 content and <5% modal biotite content (Donchak *et al.*, 2007). Its emplacement age is Lower Triassic (~247 Ma, Donchak *et al.*, 2007). During the same period, magmatism occurred simultaneously along a broad (70-90 km) belt that stretched for 300 km from northeast to southwest (Figure 1b). This magmatic belt is considered to be associated with an Early-Mid Triassic Andean-type subduction zone, which was possibly subjected to asymmetric (counterclockwise) rollback in the Upper Triassic (Li *et al.*, 2012b). Younger (230-210 Ma) granitoids are aligned along a N-S belt farther east (Figure 1b).

Following the hike in Girraween National Park, we will return to the New England Highway and continue to the Queensland – New South Wales border. The day will be concluded in the town of Tenterfield (~850 m a.s.l.), approximately ~20 km south of the state border (29 km from the last stop).

Day 2

The second day begins with a long drive south along the high altitude New England highway (~1000 m a.s.l), via the town of Glen Innes (Figure 3). The majority of the outcrops between Tenterfield and Glen Innes are late Permian to Triassic I-type granitoids and volcanic rocks. South of Glen Innes the dominant rocks are Cenozoic basalts.

Stop 2.1. Tobermory Monzogranite

GPS coordinates: -30.196364°/152.010624°

Distance from Tenterfield: 172 km

The Tobermory Monzogranite is part of the Hillgrove Plutonic Suite, which is made of early Permian S-type granitoids. A recent U-Pb SHRIMP zircon age from this pluton (from a locality ~2 km north of this stop) has yielded an age of 292.4±2.9 Ma (Rosenbaum *et al.*, 2012).

The rock is deformed with a steeply dipping fabric striking NW-SE. This orientation is parallel to a sharp linear morphology expressed as a valley, which can be followed for a distance of ~3.5 km (Figure 5a). Elsewhere in the pluton, deformation is much weaker, suggesting that the fabric recognised here is related to a localised shear zone.

Regionally, the Tobermory Monzogranite is part of a series of plutons of similar age, which form a N-S belt of S-type granitoids (Figure 1b). This belt is supposedly situated in the eastern limb of the Manning Orocline, while the western limb is represented by the Bundarra Granite (Rosenbaum, 2012). Our recent geochronological ages confirm the suggestion that these granitic plutons were emplaced simultaneously (Rosenbaum *et al.*, 2012). However, the nature and geochemistry of magmatism is somewhat different, with the Bundarra Granite being more voluminous, coarse-grained and richer in SiO_2 (Shaw and Flood, 1981).

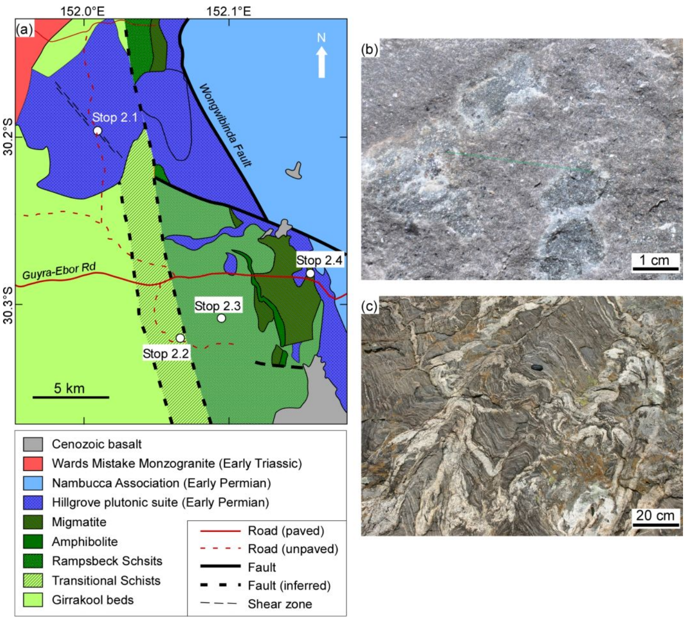
Stop 2.2. Girrakool beds (transitional schists)

GPS coordinates: -30.320311°/152.065682°

Distance from previous stop: 19 km

The early Permian S-type Hillgrove plutons, such as the Tobermory Monzogranite visited in the previous stop, were intruded into Carboniferous metasedimentary rocks of the Tablelands Complex, which are here represented by the Girrakool beds. The rocks were deposited in a subduction complex, and metamorphosed under variable metamorphic conditions. As we move eastward in the next few stops (stop 2.2 to stop 2.5), we will see a progressive increase in the metamorphic grade, from the very low grade Girrakool beds in this locality, to high-grade amphibolite-facies rocks and migmatites (Figure 5a). This metamorphic complex is normally referred as the Wongwibinda metamorphic complex (Binns, 1966; Stephenson and Hensel, 1982; Danis *et al.*, 2010; Craven *et al.*, 2012).

Figure 5. Geology of the Wongwibinda metamorphic complex.



(a) Geological map of the Wongwibinda metamorphic complex (modified after Danis *et al.*, 2010). (b) Large cordierite poikiloblasts in the Rampsbeck schists (Stop 2.3) (photo courtesy of Steve Craven); (c) Partial melting and folding in migmatites (Stop 2.4) (photo courtesy of Nathan Daczko).

The Girrakool beds in this locality are low-grade metamudstones and greywackes, located at the western margin of the metamorphic complex. The mudstones have been reconstituted to phyllites containing quartz, plagioclase, biotite, muscovite and minor graphite. The greywackes have large size clasts of primary sedimentary quartz and feldspar in a recrystallised matrix. Bedding and graded bedding are preserved. The rocks are moderately deformed with slaty cleavage developed, especially in the mudstone. Refraction of the structural fabric can be recognised, and most likely reflect primary graded bedding. Steeply plunging folds are also recognised.

Stop 2.3. Rampsbeck schists

GPS coordinates -30.308807°/152.093561°

Distance from previous stop: 4.3 km

Access to this outcrop involves walking northward from Lynoch driveway for a distance of 1.6 km and requires permission from the property owners. The protolith of these rocks is considered to be the Girrakool beds, but in this locality the sedimentary grains are completely recrystallised to a biotite-quartz-plagioclase schist with large (up to 2 cm) poikiloblasts of cordierite, k-feldspar and muscovite ± garnet (Figure 5b). Inclusions within poikiloblasts are randomly oriented and dominated by biotite and quartz. Geothermobarometry analysis on the assemblage grt-bt-plag-kfs-qtz yielded pressure-temperature estimates of 0.42±0.12 GPa and 647±36°C (Danis *et al.*, 2010).

Stop 2.4. Migmatites

GPS coordinates: -30.281746°/152.155289°

Distance from previous stop: 19.5 km

At this locality, we are situated in the highest-grade of the metamorphic complex, with rocks exhibiting amphibolite-facies conditions and migmatisation (Figure 5c). The migmatites comprise quartz, biotite, plagioclase, microcline and garnet ± cordierite and are characterised by abundant melanosome layers and folded leucosomes (Danis *et al.*, 2010). The directly calibrated pressure-temperature estimates have relatively large error bars (0.46±0.45 GPa and 764±106°C, Danis *et al.*, 2010), but demonstrate increasing temperature conditions relative to the Rampsbeck schists. Pseudosections from high-grade rocks indicate temperatures of ~660°C and pressures of <0.35 GPa (Craven *et al.*, 2012).

Close to this locality, the high-temperature metamorphic rocks have been intruded by another pluton of the Hillgrove suite, the Abroi Granodiorite. Recent geochronological data by Craven *et al.* (2012) (U–Th–Pb monazite age) shows that peak metamorphism occurred at 296.8±1.5 Ma, which is ~6-7 Ma prior to the emplacement of Abroi Granodiorite (~290 Ma, Cawood *et al.*, 2011a; Craven *et al.*, 2012). These results indicate that it is not the local heating associated with magmatism that was responsible for high-temperature metamorphism. On the contrary, S-type magmatism occurred after peak metamorphism, and was most likely associated with melting of the subduction complex rocks. The observations from the Wongwibinda metamorphic complex, together with time constraints on the emplacement of S-type granitoids, indicate that during the early Permian (300-290 Ma), these rocks were situated in an anomalously hot tectonic setting. In the context of a convergent margin, this tectonic environment would most likely be the backarc region, which is typically characterised by anomalously high heat flow (e.g. Hyndman *et al.*, 2005). The coincidence of regional high-temperature metamorphism, crustal melting, bimodal volcanism and the development of rift-related sedimentary basins (Korsch *et al.*, 2009, and Stop 3.1) supports the idea that early Permian tectonics in the New England Orogen was primarily controlled by backarc extension.

Stop 2.5. Ebor Falls

GPS coordinates: -30.403008°/152.342914°

Distance from previous stop: 29.5 km

After crossing the Wongwibinda Fault (5a), we travel east through low-grade metasedimentary rocks, before entering into a large area covered by Cenozoic basalts.

The spectacular Ebor Falls exposes a thick section of basaltic lava flow, overlying early Permian metasedimentary rocks of the Nambucca Association. Structurally, we are located in the southern continuation of a large N-S trending strike-slip fault, the Demon Fault (Figure 1b). Farther north, the Demon Fault has a clear geological and morphological expression for approximately 150 km, but in this area and farther south it becomes less clear. The Demon Fault has a post-Triassic dextral strike-slip movement of 25-30 km (McPhie and Fergusson, 1983).

The basalts are part of a central volcano (Ebor Volcano) dated at 19-20 Ma (Ashley *et al.*, 1995). The lava flow has a substantial thickness of up to 400 m and covers an area of 480 km^2. Rocks are mostly olivine- and quartz-normative tholeiitic basalts, with some variations to alkaline and transitional basalts (Ashley *et al.*, 1995).

The Ebor Volcano is one in a series of Oligocene to Miocene central volcanoes in eastern Australia. These volcanoes become progressively younger from north to south, suggesting that they correspond to a hotspot track above a fixed mantle plume. The motion path of the Australian plate relative to this hotspot has recently been unravelled by ^{40}Ar/^{39}Ar geochronology of the central volcanoes (Knesel *et al.*, 2008), showing a pronounced deceleration in the motion of Australia at 26-23 Ma.

Stop 2.6. Newell Falls - Dorrigo Mountain Complex

GPS coordinates: -30.393299°/152.745943°

Distance from previous stop: 56 km

The winding road from Ebor Falls descends eastward through Cenozoic basalts, before reaching the underlying early Permian rocks. The Dorrigo Mountain Complex is a heterogeneous igneous body comprising fine-grained diorite and granite, which seems to intrude the basal layers of the early Permian Nambucca succession. Recent SHRIMP U-Pb zircon dating from this locality yielded an age of 295.5±3.0 Ma (Rosenbaum *et al.*, 2012).

The igneous complex is one in a series of early Permian plutons that occur along a NW-SE belt between Glen Innes and Coffs Harbour (Figure 1b). This belt is the eastern continuation of the Hillgrove plutonic suite farther west, with both segments emplaced simultaneously at 295-290 Ma (Rosenbaum *et al.*, 2012). The granitic belt is curved around the area of the Henry River Granite (Figure 1b), thus defining another strong bend in the structural grain of the southern New England Orogen. This bend, referred to by Rosenbaum *et al.* (2012) as the Nambucca Orocline, could be the internal hinge of the Texas Orocline (Figure 2b). Farther to the southeast, the orocline is represented by the Hastings Orocline (Glen and Roberts, 2012).

Travelling east from this locality, the narrow road continues to meander through a lush rainforest. The rocks immediately adjacent to the Dorrigo Mountain Complex, the McGraths Hump basalts, are early Permian in age, and are possibly part of the same igneous complex. These rocks are equivalent in age and nature to other bimodal volcanic rocks within the early Permian Nambucca Block, such as the Petroi Basalt (Asthana and Leitch, 1985) and Halls Creek Volcanics (Leitch, 1988; Cawood *et al.*, 2011a). The age of the latter has recently been dated as 292.6±2.0 (Cawood *et al.*, 2011a). The importance of these rocks is that they all occur close to the base of the early Permian sedimentary succession of the Nambucca Block, and therefore provide an age constraints for basin development. The collective observations on bimodal volcanism, high-temperature metamorphism (Stops 2.4 and 2.5), crustal melting (Stop 2.1) and rift-related sedimentation (Stop 3.1), further support a backarc extensional environment during the early Permian.

The second day concludes in the coastal town of Nambucca Heads, 57 km from the last stop.

Day 3

This day involves relatively short driving distances and benefits from spectacular outcrops in the coastal headlands between Nambucca Heads and Port Macquarie (Figure 3). The first outcrop is located in Nambucca Heads, within a walking distance from the town centre.

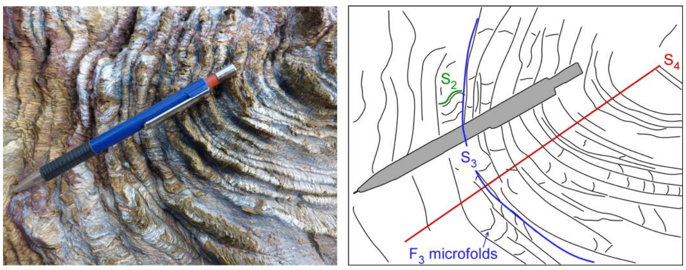
Stop 3.1. Nambucca Head

GPS -30.647226°/ 153.019757°

The headland at the southern end of Shelly Beach provides an excellent exposure of deformed early Permian rocks of the Nambucca Block. The protolith is a siltstone sequence that was supposedly deposited in an extensional rift basin (Korsch *et al.*, 2009), considered to be part of the larger Barnard Basin (Leitch, 1988). In this locality, the rocks are metamorphosed at low greenschist facies conditions and show multiple overprinting crenulation fabrics (Leitch, 1978; Johnston *et al.*, 2002).

Four generations of structural fabrics can be recognised in this outcrop. The dominant fabric element is defined by the layering of quartz and mica domains (Figure 6) and is generally oriented ~N-S. The differentiation to quartz-mica domains is a characteristic feature of intense crenulation cleavage (Gray, 1977), indicating that the dominant fabric must be an overprinting fabric element (at least an S2) that was superimposed on an earlier structural fabric (S1). In thin sections, traces of S1 are locally recognised in low strain zones within the quartz-rich S2 domains. The prominent crenulations in the quartz-mica domains are F3 folds, and this crenulation cleavage (S3) is further folded around F4 folds (Figure 6).

Figure 6. Nambucca Heads.



Overprinting deformation fabrics in early Permian rocks, Nambucca Heads (Stop 3.1). Note that traces of S1 can only be recognised in low strain zones under the microscope.

The tectonic context of the multiple deformation episodes in the Nambucca Block, and relationship with the process of oroclinal bending, are not entirely understood. Offler and Foster (2008) among others have argued that the earlier E-W structural fabric (S1; better preserved elsewhere within the Nambucca Block), formed during southward indentation of the Coffs Harbour Block. Southward indentation, according to Offler and Foster (2008), was responsible for the formation of the Coffs Harbour Orocline. Offler and Foster (2008) have used Ar-Ar dating to constrain the timing of S1 fabric development at 264-260 Ma, arguing that these ages constrain the timing of oroclinal formation. This model assumes that the Nambucca Block is autochthonous and that oroclinal bending was primarily controlled by a N-S tectonic transport. Alternatively, it is possible that the Nambucca Block is a displaced terrane that was subjected to vertical axis block rotations during oroclinal bending (e.g. Collins *et al.*, 1993). In my opinion, this is a reasonable proposition, given the fact that map-scale curvatures are recognised in the contemporaneous belt of early Permian granitoids. Whether the observed overprinting relationships between structural fabrics represent a progressive rotation of the whole Nambucca block, or changes in the direction of maximum shortening, remains an open question.

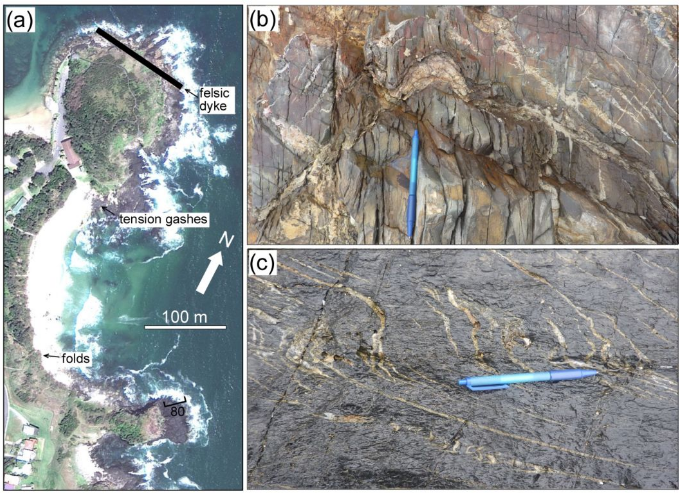
Stop 3.2. Scotts Head

GPS -30.745117°/152.998174°

Distance from previous stop: 31 km

The next headland south of Nambucca Heads also exposes early Permian rocks of the Nambucca succession. Rocks are low-grade fine-grained carbonaceous metasiltstones, metasandstones and calcareous metasiltstones, intruded by younger (Triassic?) felsic dykes (Figure 7a). The structure has been described by Johnston *et al.* (2002), who recognised microscopic isoclinal F1 folds (in thin sections), overprinted by steeply plunging F2 folds (Figure 7b) and associated NE-SW axial plane cleavage. Later structures include en echelon tension gashes (Figure 7c) and kink bands.

Figure 7. Geology of Scotts Head (Stop 3.2).



(a) Areal (Google Earth) image of the rock platforms in Scotts Head and indications of geological features. (b) Mesoscopic folding with axial planes parallel to the dominant structural fabric. (c) Dextral sense of shear indicated by quartz-filled tension veins.

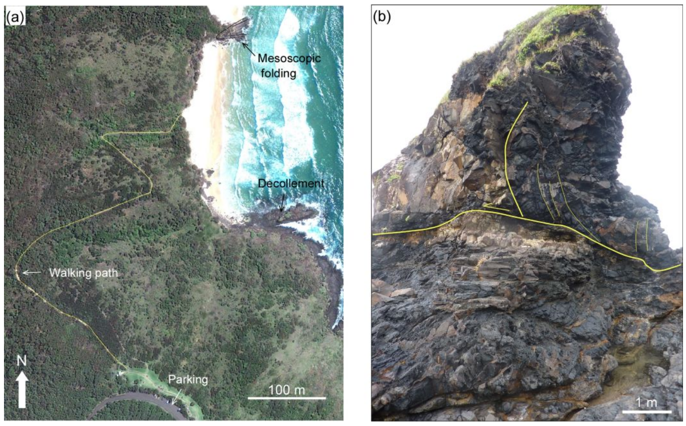
Stop 3.3. Smoky Cape

GPS: -30.916241°/153.086200°

Distance from previous stop: 51.6 km

A short (800 m) hike from the car park brings us to beach outcrops (Figure 8a). The rocks in this locality also belong to the early Permian Nambucca Association, and are similar to the metasedimentary succession in Scotts Head (Stop 3.2). However, the outcrop provides an opportunity to better understand the style of folding in the Nambucca Block. The outcrop at the northern edge of the beach shows steeply dipping beds and a mesoscopic fold that is moderately plunging to the NE. The NE-SW axial plane orientation is similar to the axial plane of the F2 folds in Scotts Head. In the southern end of the beach, a recumbent fold seems to be detached from the underlying rocks by sub-horizontal thrust décollements (Figure 8b).

Figure 8. Geology of Smoky Cape (Stop 3.3).



(a) Areal (Google Earth) image showing the coastal outcrops and the trace of the walking trail from the parking area. (b) Mesoscopic folding on top of a sub-horizontal thrust décollement.

Stop 3.4. Crescent Head

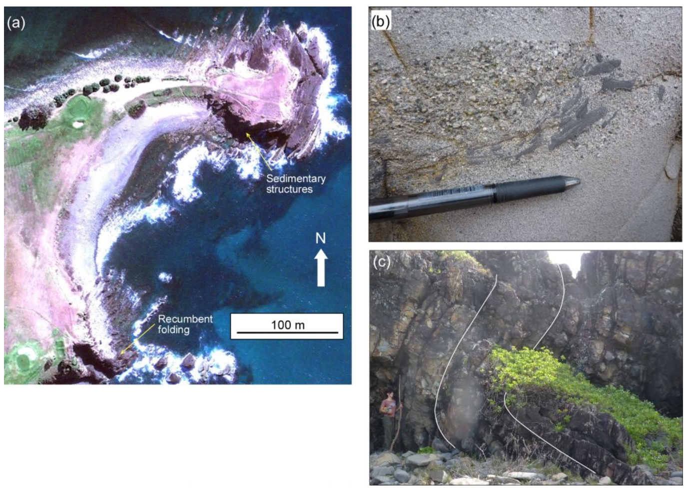
Distance from previous stop: 47.5 km

Stop 3.4.1. Sedimentary structures

GPS: -31.186872°/152.983673°

Rocks in this outcrop belong to the Kempsey beds, which is an early Permian succession of turbiditic lithic sandstone, siltstone, diamictite and conglomerate (Runnegar, 1970; Roberts *et al.*, 1995b). Rocks are shallowly dipping to the NW, and are characterised by sandstones with a coarser grained basal interval that has a variety of gravel-sized clasts (Figure 9b). Grey mudstone clasts display current imbrication, in which the shallow dip direction of the inclined clasts points upstream.

Figure 9. Crescent Head.



(a) Geology of Crescent Head (Stop 3.4). (a) Areal (Google Earth) image of Crescent Head. (b) Sedimentary structures showing a coarse grained sandstone layer and imbricated mudstone clasts (Stop 3.4.1). (c) Recumbent fold with a very high interlimb angle (Stop 3.4.2).

Stop 3.4.2. Recumbent folding

GPS: -31.188876°/152.982350°

South of the beach, there are variations in the bedding orientations and mesoscopic recumbent folds (Figure 9c). Similarly to the folds in Smoky Cape (Stop 3.3), these folds are underlain by horizontal décollements. Asymmetric en echelon tension gashes suggest sinistral shearing.

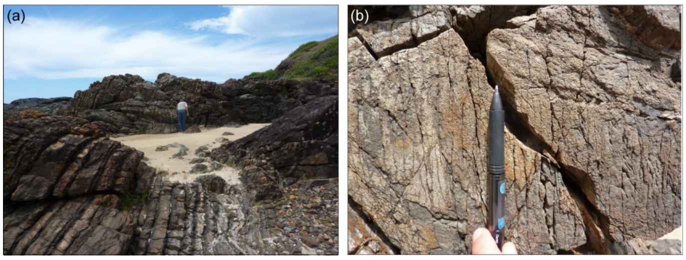
Stop 3.5. Racecourse Head

GPS: -31.251065°/152.968302°

Distance from previous stop: 9 km

The last stop today is a well-preserved package of alternating siliceous metasandstone and metasiltstone (Figure 10a). The rocks, which belong to the Upper Devonian Touchwood Fm (Roberts *et al.*, 1995b), are part of the northern Hastings Block, and were deposited in the forearc region of the Devonian-Carboniferous subduction zone. The stratigraphy is supposedly correlative to the Tamworth Belt (Roberts *et al.*, 1995b; Glen and Roberts, 2012) (Figure 1b). In this locality, rocks are steeply dipping and oriented E-W. A spaced subvertical structural fabric, oriented ~N-S (perpendicular to bedding), is also recognised (Figure 10b).

Figure 10. The Hastings Block in Racecourse Head (Stop 3.5).



(a) Steeply dipping strata of alternating siliceous metasandstone and metasiltstone layers. (b) Spaced subvertical ~N-S structural fabric perpendicular to bedding.

The Hastings Block was possibly displaced and rotated to its current position during the formation of the Manning Orocline. Palaeomagnetic data indicate that the Hastings Block was subjected to a large degree of block rotation, but whether this rotation was clockwise (Schmidt *et al.*, 1994) or counterclockwise (Klootwijk, 2009) is debated. Cawood *et al.* (2011b) recently concluded that existing palaeomagnetic data are insufficient to support a unique kinematic reconstruction of the Hastings Block and Manning Orocline. Nevertheless, the authors suggested a preferred model, which involves oroclinal bending accompanied by large displacement sinistral shearing. An alternative model involving much smaller displacements and no shearing has recently been proposed by Glen and Roberts (2012).

The day concludes in the town of Port Macquarie, a distance of 72.5 km from the last stop.

Day 4

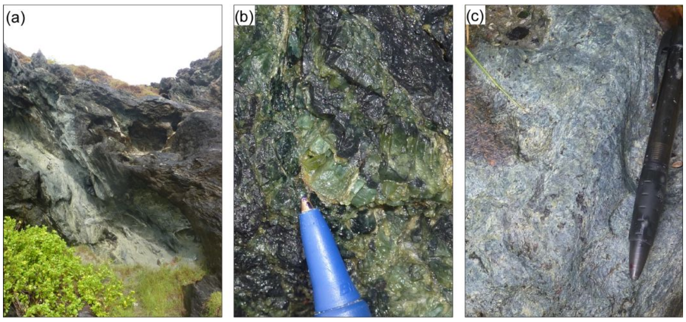
Stop 4.1. Port Macquarie serpentinites and high-pressure rocks

GPS: -31.435876°/152.925736°

The coast of Port Macquarie provides a beautiful exposure of mafic and ultramafic rocks, including serpentinites and high-pressure assemblages. The origin of these rocks, possibly associated with an ophiolitic mélange (Aitchison *et al.*, 1994), is Cambrian-Ordovician (Aitchison and Ireland, 1995; Fukui *et al.*, 1995). A similar lithological assemblage occurs in a number of other localities throughout the southern New England Orogen (Figure 1b), most noticeably along the length of the Peel-Manning Fault System (Figure 1b). Farther south, the southwards continuation of the Peel-Manning Fault System is unclear, but patches of serpentinites are found in the area of the Manning Orocline and around the Hastings Block (Lennox and Offler, 2009). Therefore, it is possible that the same serpentinite belt that crops out along the Peel-Manning Fault System is folded around the Manning Orocline and the Hastings bend (i.e. Nambucca Orocline, Figure 2b) before reappearing in Port Macquarie. It is noted, however, that the correlation between the different serpentinite outcrops is not straightforward. For example, some of the ophiolitic-related rocks on the western side of the Hastings Block are probably Silurian-Devonian (Aitchison *et al.*, 1994), i.e., considerably younger than the Cambrian-Ordovician Port Macquarie rocks.

A detailed description of the geology of Port Macquarie has been provided by Och *et al.* (2003; 2007). The rocks are characterised by blocks of massive serpentinites and mafic rocks surrounded by a serpentinite schist (Figure 11a). The serpentinite assemblage in both blocks and matrix is lizardite and chrysotile (Och *et al.*, 2007). Some of the blocks preserve cumulative textures and show primary magmatic layering. A protolith of harzburgite, lherzolite and orthopyroxenite has been suggested for the serpentinite blocks (Och *et al.*, 2007).

Figure 11. Serpentinites and high-pressure rocks in Port Macquarie (for detailed geological maps see Och *et al.*, 2007).



(a) Sheared serpentinite matrix surrounding blocks of massive serpentinite and high-pressure rocks. (b) Coarse-grained tremolite crystals; (c) A blueschist facies assemblage of glaucophane-phengite schist.

High-pressure rocks occur as blocks within a metamorphic mélange, and include retrogressed eclogite- and blueschist-facies metamorphic assemblages of almandine, omphacite, ±lawsonite, ±glaucophane and ±quartz (Och *et al.*, 2003). Blocks of tremolite marble and omphacitite are also found (11b). Some layers of garnet-omphacite-lawsonite-quartz are intercalated with phengite glaucophane schist (Figure 11c). The matrix is chlorite-actinolite schist, most likely associated with metasomatic interactions between the mafic and ultramafic rocks.

Stop 4.2. Black Head

GPS: -32.070783°/152.547435°

Distance from previous stop: 102 km

Travelling south from Part Macquarie, we pass through Triassic strata of the Lorne Basin and Upper Triassic granitoids (Figure 1b) before approaching the Devonian-Carboniferous rocks of the Myall Block. Palaeomagnetic data indicate that the Myall Block, which is part of the forearc basin (Tamworth Belt), underwent 120° of counterclockwise rotations (Geeve *et al.*, 2002). Such rotations around vertical axis could possibly be attributed to oroclinal bending in the Manning Orocline and the associated displacement of the Myall Block to a position at the southeast limb of the orocline.

The rocks in this locality belong to the Devonian Bundock beds. They are moderately dipping to the ENE. A fault striking ~200° is recognised, with kinematic indicators (tension gashes and minor folds) suggesting a sinistral sense of movement.

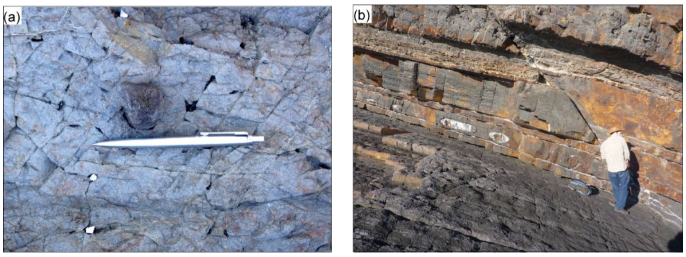
Stop 4.3. The Myall Block near Smiths Lake

GPS: -32.378914°/152.532545°

Distance from previous stop: 50 km

A series of headlands between Charlotte Head and Seal Rocks offer fine exposures of the Carboniferous forearc-related sedimentary package. At this locality, south of Danger Point, sandstone, siltstone and clay layers of the Wootton Beds are shallowly dipping (20°-25°) to the west, forming extensive dip slopes. Clasts of volcanic rocks (e.g. scoria) are found within the fine-grained clastic rocks (Figure 12a), indicating proximity to the volcanic source. Deformation is relatively weak, but a few E-verging reverse faults can be recognised (Figure 12b).

Figure 12. The Myall Block near Sandbar (Stop 4.3).



(a) Clast of scoria within siltstone (photo courtesy of Uri Shaanan); (b) minor E-verging reverse fault.

Stop 4.4. Nerong Volcanics in Anna Bay

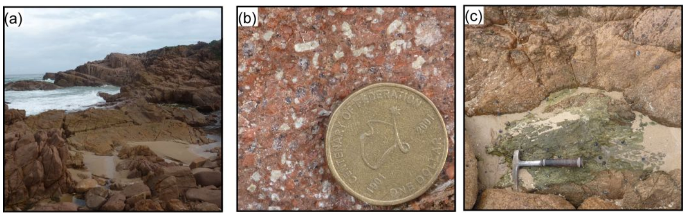
GPS: -32.787639°/152.081216°

Distance from previous stop: 134 km

The Devonian-Carboniferous magmatic arc is rarely exposed in the southern New England Orogen, and is mostly inferred from provenance studies of the forearc basin (Tamworth Belt and correlative block). This locality provides the only real exposure of genuine Carboniferous volcanic rocks (Figure 13a) that possibly represent the actual volcanic arc (Buck, 1988). Based on U-Pb geochronology, the age of the Nerong Volcanics has been constrained at 338.6±3.8 Ma (Roberts *et al.*, 1995a).

The rocks are ignimbrites, containing phenocrysts of quartz, K-feldspar and large weathered lithic fragments within a red groundmass (Figure 13b). Flow textures are not common, but there is evidence for the incorporation of large (~1 m) mafic blocks, metamorphosed to epidote-chlorite greenschist, and surrounded by ignimbrite (Figure 13c).

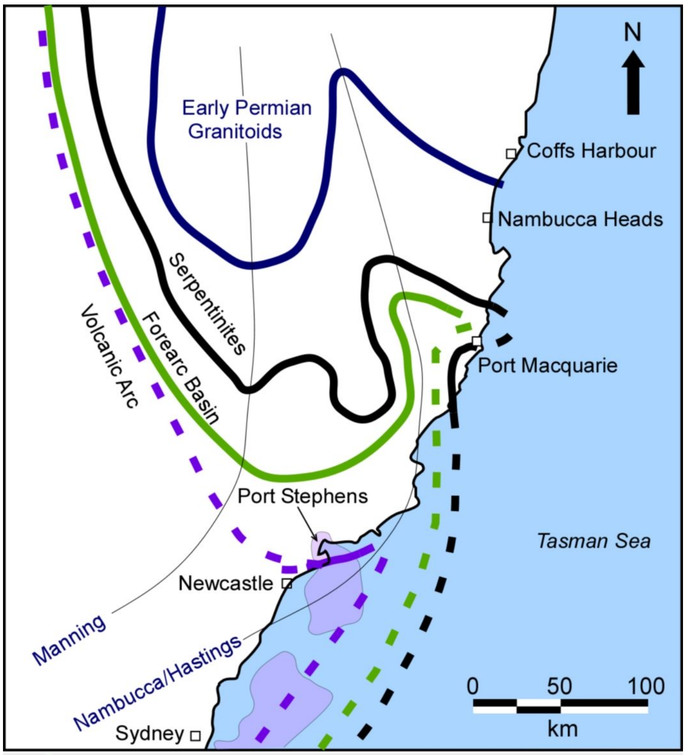
Figure 13. The Nerong Volcanics in Anna Bay.



(a) Exposures of ignimbrite slabs. (b) Fragments of quartz, K-feldspar and weathered lithic fragments in a red groundmass. (c) A block of epidote-chlorite metabasalt surrounded by ignimbrite.

The exposure of the Carboniferous volcanic arc in this locality is consistent with the large-scale oroclinal structure (Figure 14). As recently pointed out by Glen and Roberts (2012), data from petroleum exploration suggest that pre-Permian volcanic rocks occur along a NNE-SSW belt offshore Sydney. The volcanic arc is exposed in the area around Port Stephens (i.e. Anna Bay), but does not continue northward into the Myall and Hastings Block. Rather, it is most likely underthrust beneath the Tamworth Belt farther to the northwest (Figure 14). The gentle curvature and sharp cusp in the volcanic arc were inferred to reflect the external hinge of the southern oroclines (Manning and Nambucca/Hastings), which progressively decrease their amplitudes from north to south (Glen and Roberts, 2012, Figure 14).

Figure 14. Tectonic interpretation of the Manning/Nambucca oroclines.



An orogenic-scale interpretation of the Manning and Nambucca (Hastings) oroclines, highlighting the reason why Carboniferous volcanic arc rocks (Nerong Volcanics) are exposed in Anna Bay. The inferred exposure of arc-related volcanic rocks offshore Sydney is after Glen and Roberts (2012). Onshore, Carboniferous arc volcanic rocks are interpreted to be underthrust beneath the forearc basin rocks.

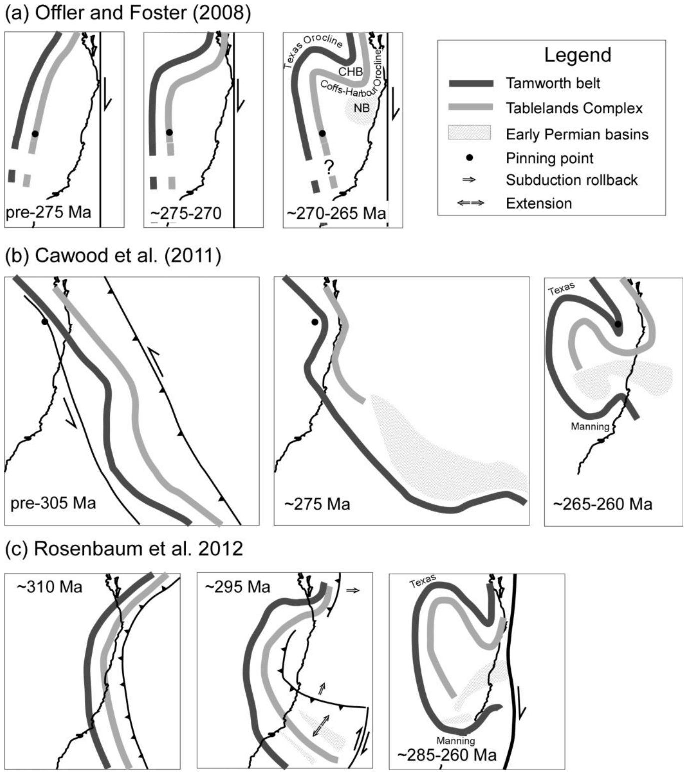
The fieldtrip concludes in Sydney Airport, a distance of 214 km from the last stop.

Discussion and conclusions

The collective set of observations from the southern New England Orogen is consistent with a geometrical model of a regional ear-shaped structure hundreds of km across. This geometry involves four major oroclines (Texas, Coffs Harbour, Manning and Nambucca/Hastings), and is different than previous interpretations, in which a simpler structure of two oroclines (Murray *et al.*, 1987; Offler and Foster, 2008) or three oroclines (Cawood and Leitch, 1985; Korsch and Harrington, 1987) was assumed. Although the fieldtrip did not permit careful observations of all the hinge zones, it has provided an introduction to the major geological components that define the oroclinal structure, in particular the Devonian-Carboniferous convergent margin (subduction complex, forearc basin and magmatic arc), the serpentinite belt, and the early Permian belt of granitoids.

Tectonic reconstructions of the New England oroclines are still oversimplified and require further structural, geochronological and palaeomagnetic constraints. Three alternative schematic reconstructions are presented in Figure 15. The reconstruction of Offler and Foster (2008) explains the formation of the Texas-Coffs Harbour oroclines and assumes that the southern oroclines do not exist. The proposed model (Figure 15a) considers that the oroclines were generated by an offshore dextral strike-slip fault. Movement on such a transform fault was supposedly responsible for a progressive bending of the earlier convergent margin, which was pinned in the area around Tamworth. The model involves large ~N-S displacements, particularly in the area of the Coffs Harbour Orocline, with the supposed indentation of the Coffs Harbour Block onto the Nambucca Block. Offler and Foster (2008) concluded that the earlier E-W deformation in the Nambucca Block, dated at ~270-265 Ma, was directly related to this indentation, thus marking the last stage of oroclinal bending.

Figure 15. Alternative tectonic models for the formation of the New England oroclines.



(a) Oroclinal bending generated by an offshore dextral strike-slip faulting (modified after Offler and Foster, 2008). CHB, Coffs Harbour Block; NB, Nambucca Block. Note that the model only accounts for the formation of the Texas and Coffs Harbour oroclines. (b) Formation of the oroclines by buckling within a zone of sinistral transpression (modified after Cawood *et al.*, 2011b). (c) A schematic model showing the progressive curvature of the plate boundary during subduction rollback, followed by further bending in a strike-slip setting (modified after Rosenbaum *et al.*, 2012).

There are a number of problems with the model of Offler and Foster (2008). Firstly, their dextral strike-slip model does not explain the curvature of the southernmost oroclines (Manning and Nambucca oroclines). Secondly, as recently pointed out by Li *et al.* (2012a), the predicted strain from the strike-slip model (~500 km displacement and ~50% shortening) is considerably larger than the observed strain in the area of the Texas Orocline. Thirdly, Cawood *et al.* (2011b) have recently demonstrated that the dextral strike-slip model is not consistent with available palaeomagnetic data.

The model by Cawood *et al.* (2011b) (Figure 15b) is based on a synthesis of available but limited palaeomagnetic data from the southern New England Orogen. However, this dataset is insufficient for the construction of an unequivocal model (Pisarevsky *et al.*, 2010), and therefore suffers from large uncertainties. The major assumption by Cawood *et al.* (2011b) is that oroclinal bending was governed by sinistral strike-slip tectonics (Figure 15b), which involved a very large northward displacement (~1600 km) of the Hastings Block. Glen and Roberts (2012) have recently argued that such large-scale displacements are unlikely. In my opinion, the major shortcoming of Cawood *et al.*'s (2011b) model is the fact that they attributed the entire process of oroclinal bending to transpressional tectonics, without explaining the plethora of evidence for syn-oroclinal extension.

An alternative model (Rosenbaum *et al.*, 2012) is based on the assumption that subduction rollback and back-arc extension have played a primary role in controlling the process of oroclinal bending (Figure 15c). This process is similar to many modern examples, in which tight orogenic curvatures were obtained by along-strike variations in the rate of subduction rollback and slab segmentation (Barker, 2001; Schellart *et al.*, 2002; Rosenbaum and Lister, 2004; Rosenbaum and Mo, 2011). In the New England Orogen, it is possible that earlier rollback-related curvatures, as well as other irregularities in shape of the convergent margin (e.g. Glen and Roberts, 2012; Li *et al.*, 2012a), were subjected to further tightening during subsequent events of contraction and transpression.

The key to understanding the geodynamics of the New England Orocline, in my opinion, is the information from early Permian rocks. Evidence for early Permian extensional tectonics, high-temperature metamorphism and crustal melting supports the suggestion that this period was characterised by a retreating, perhaps Mediterranean or SW Pacific style, subduction boundary.

Acknowledgements

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