Cenozoic deformation in the Otway Basin, southern Australian margin: implications for the origin and nature of post-breakup compression at rifted margins

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ABSTRACT

There is growing recognition that pulses of compressive tectonic structuring punctuate the post-breakup subsidence histories of many ‘passive’ rifted continental margins. In order to obtain new insights into the nature and origin of compression at passive margins we have conducted a comprehensive analysis of the post-breakup (<43 Ma) deformation history of the offshore Otway Basin, southern Australian margin, using a regional seismic database tied to multiple wells. Through mapping of a number of regional intra-Cenozoic unconformities we have determined growth chronologies for a number of major anticlinal structures, most of which are ~NE-SW-trending folds that developed during mild inversion of syn-rift normal faults or through buckling of the post-rift succession. These chronologies are supplemented by onshore structural evidence and by thermochronological data from key wells. Whilst our analysis confirms the occurrence of a well-documented pulse of late Miocene–early Pliocene compression, post-breakup deformation is not restricted to this time interval. We highlight the growth of a number of structures during the mid-late Eocene and the Oligocene-early

Miocene, with evidence for considerable temporal and spatial migration of strain within the basin. Our results indicate a long-lived ~NW-SE maximum horizontal stress orientation since the mid-late Eocene, consistent with contemporary stress observations but at variance with previous suggestions that this stress orientation was initiated in the late Miocene by increased coupling of the Australian-Pacific plate boundary. We attribute the observed record of deformation to a compressional intraplate stress field, coupled to the progressive evolution of the boundaries of the Indo-Australian Plate, ensuring that this margin has been subject to ongoing compressional forcing since mid-Eocene breakup. Our results indicate that compressional deformation at passive margins may be more common than is generally assumed, and that passive margin basins with evidence for protracted post-breakup deformation histories can provide useful natural laboratories for obtaining improved understanding of the evolution of intraplate stress fields over geological timescales.

INTRODUCTION

In recent years, seismic exploration of rifted continental margins has provided growing evidence that the post-rift evolution of some margins is not characterized solely by tectonic quiescence and dominantly extensional stress regimes, as implicit in many models of 'passive' margin evolution (Bond & Kominz, 1988). In fact, the post-rift subsidence histories of many passive continental margins worldwide have been punctuated by pulses of localised compressional shortening (i.e. tectonic inversion; Williams *et al.* 1989; Turner & Williams, 2004) that are often accompanied by regional uplift, producing long-wavelength (>200-500 km) low-angle (≤5°) unconformities (Doré *et al.,* 2002; Praeg *et al.,* 2005; Johnson *et al.,* 2008). Margins that provide evidence for post-rift compression include: the NW European Atlantic margin (Boldreel & Anderson, 1998; Vågnes *et al.,* 1998; Lundin & Doré, 2002; Davies *et al.,* 2004; Stoker *et al.,* 2005; Williams *et al.,* 2005; Doré *et al.,* 2008; Hillis *et al.,* 2008*a*; Holford *et al.*, 2008, 2010*a*; Tuitt *et al.*, 2010); the Iberian margin (Peron-Pinvidic *et al.*, 2008; Vasquez *et al.*, 2008; Pereira *et al.*, 2011); the eastern US margin (Kulpecz *et al.*, 2009); the NE Brazilian margin (Cobbold *et al.*, 2001, 2007, 2010); the western African (Angolan) margin (Hudec & Jackson, 2002); and the Australian NW Shelf (Harrowfield & Keep, 2005; Hillis *et al.*, 2008*b*). Constraining the timing, distribution and ultimately the origin of post-rift compression has important economic implications, as compressional folding of syn-rift and post-rift sedimentary successions can produce significant hydrocarbon traps (e.g. the Ormen Lange Dome on the mid-Norwegian margin; Vågnes *et al.*, 1998). Moreover, the near-continuous history of post-rift sedimentary deposition at many margins provides them with the potential capability of serving as sensitive recorders of geodynamic events and processes. Careful examination of the timing and distribution of post-breakup compressional structures and unconformities at passive margins may thus yield important insights to generic problems like the evolution of intraplate stress fields (Zoback & Zoback, 2007) or the impact of mantle flow on surface topography (Braun, 2010).

This paper examines the record of post-breakup compression in the offshore Otway Basin of the southern Australian margin. The Otway Basin is one of several depocentres that formed during Cretaceous-Palaeogene continental breakup between Antarctica and Australia (Norvick & Smith, 2001). Along this margin, compressional structures have been most commonly identified within the post-rift Cenozoic successions of the Otway, Gippsland and Bass basins (Hill *et al.*, 1995; Perineck & Cockshell, 1995; Dickinson *et al.*, 2001; Hillis *et al.*, 2008*b*), and many significant producing oil and gas fields in these basins are hosted within Cenozoic anticlines (Bernecker *et al.*, 2003; Holford *et al.*, 2011*a*). Similar Cenozoic anticlines occur within the Torquay sub-basin (Holford *et al.*, 2011*b*) but drilling has shown that they do not contain significant quantities of hydrocarbons (Trupp *et al.*, 1994), implying that the relative timing of compressional deformation with respect to hydrocarbon generation and charge is a key factor in defining prospectivity along the margin (Duddy, 1997; Holford *et al.*, 2010*b*).

Previous studies in the Otway Basin have identified major periods of compressional deformation and resultant fold growth during the mid-Eocene (Duddy *et al.*, 2003) and late Miocene-Pliocene (Dickinson *et al.*, 2001, 2002). It is generally accepted that Cenozoic deformation of the southern Australian margin (which continues to the present-day as evidenced by elevated levels of intraplate seismicity and seismogenic strain rates of up to $\sim10^{16}$ s^{-1} (Célérier *et al.*, 2005; Sandiford & Egholm, 2008; Holford *et al.*, 2011*c*) is controlled to first-order by plate boundary forces (Fig. 1) (Sandiford *et al.*, 2004; Hillis *et al.*, 2008*b*; Müller *et al.*, 2012). As is common with many studies of post-rift compression at passive margins (e.g. Holford *et al.*, 2009*a*), previous investigations in the Otway Basin have often sought to identify a sole driving mechanism responsible for the major pulses of deformation, usually a far field, plate boundary tectonic event. For example, late Miocene-Pliocene fault reactivation and faulting in the Otway Basin has been variously ascribed to the contemporaneous collision of Australia's northern margin with the island arc in New Guinea (Hill *et al.*, 1995), or increased coupling of the Australian-Pacific plate boundary resulting from formation of the Southern Alps in New Zealand (Fig. 1) (Dickinson *et al.*, 2002). Whilst such events undoubtedly exert a significant control on the coeval intraplate stress field (Sandiford *et al.*, 2004), recent investigations of the timing of Cenozoic post-breakup deformation along the NW European Atlantic margin have described multiple structures with evidence for growth at various times subsequent to early-Eocene breakup, indicating that post-breakup deformation is more common than generally assumed and not necessarily restricted to discrete temporal pulses (Vagnes *et al.*, 1998; Doré *et al.*, 2008; Ritchie *et al.*, 2008; Holford *et al.*, 2009*a*).

Here we present the first comprehensive analysis of the timing of Cenozoic post-breakup (<43 Ma) shortening in the Otway Basin of the southern Australian margin, using a regional 2D seismic database calibrated to biostratigraphic data from exploration wells (Fig. 2). We have identified a number of major offshore compressional structures and where possible have correlated them with contiguous terrestrial structures identified by onshore mapping studies. Our analysis indicates that the growth of these structures, and thus the chronology of post-breakup compressional deformation in the Otway Basin, is not restricted to a narrow time interval, and there appears to have been substantial spatial and temporal migration of strain within the basin. Individual structures display a variety of ages from mid-Eocene to early Pliocene, although several periods of enhanced deformation (mid-Eocene and late Miocene) are apparent. Although some peaks in tectonic activity coincide with far-field events at plate boundaries, such as collisional events in New Guinea and New Zealand, we suggest that no single event is responsible for the record of deformation described herein. Instead, we propose that post-breakup deformation in the Otway Basin reflects an evolving compressional intraplate stress field controlled by progressive reconfigurations of the Indo-Australian plate boundaries over the past ~43 Myr, with key episodes of enhanced deformation corresponding to major reorganisations of the Indo-Australian Plate boundaries.

TECTONIC SETTING OF THE OTWAY BASIN

The Otway Basin is a large, broadly NW-SE trending extensional basin (maximum total sediment thickness ~13 km) encompassing onshore and offshore parts of South Australia (SA) and Victoria, and Tasmanian waters (Fig. 2). The basin developed following late

Jurassic-early Palaeogene rifting, breakup and eventual separation of Australia and Antarctica (Krassay *et al.,* 2004). Exploration of the basin is at a mature stage with over 200 wells drilled in the basin, resulting in several commercial gas discoveries offshore Victoria and numerous smaller onshore gas and CO_2 fields. To date, there have been no commercial oil discoveries. Most offshore exploration has been in relatively shallow water depths (<250 m), but is progressively focused on deep-water plays. Amrit-1 was drilled in 1,425 m of water in 2004, representing the deepest well drilled along the southern Australian margin yet, and acreage in the western Otway Basin released in 2012 reaches water depths of 3000 m (Totterdell, 2012).

A number of variably exposed tectonostratigraphic basement provinces underlie the Otway Basin (Fig. 2d). Much of the South Australian and western Victorian basin is underlain by terranes that form part of the Cambrian-Ordovician Delamerian fold belt (e.g. the Kanmantoo-Glenelg Zone and Grampians-Stavely Zone) or the mid-Palaeozoic Lachlan fold belt (e.g. the Stalwell Zone and Bendigo Zone) (Cayley *et al.,* 2002; Gibson *et al.*, 2011). The boundaries between these terranes strike broadly ~N to NNW. These terranes are separated from the Neoproterozic to Cambrian Selwyn-Taswegia Block, which underlies most of the Torway sub-basin and Tasmania by a major ~N-S striking boundary named the Avoca Fault (Gibson *et al.,* 2011). The Avoca Fault is interpreted to continue offshore as the Sorell Fault, a major reactivated basement structure that accommodated significant transform motion during the separation of Australia and Antarctica, which may continue into the Southern Ocean as the Tasman Fracture Zone (Gibson *et al.*, 2011, 2013). An abrupt change in the orientation of Cretaceous normal faults, from ~NW-SE to the west of the Avoca Fault to ~NE-SW to the east points to an important basement control on later rifting episodes (Miller *et al.*, 2002).

Rifting in the Otway Basin commenced in the late Jurassic-early Cretaceous (Fig. 3) first in the west then progressing to the east, resulting in several distinct grabens and half-grabens with varying geometries and orientations (Fig. 2) (Perineck & Cockshell, 1995; Krassay *et al.*, 2004). The early Cretaceous rift axis is located onshore and consists of ~W to NW trending depocentres in the west (e.g. the Robe and Penola Troughs in SA) and ~NE trending depocentres in the east (e.g. Torquay sub-basin in Victoria), with the variation in trends largely reflecting the basement heterogeneities described above (Miller *et al.*, 2002; Krassay *et al.*, 2004). Initial rift fills were dominated by carbonaceous lacustrine shales with minor interbedded sandstones and volcanics (Casterton Formation), and as the rate of extension increased into the Berriasian and Barremian syn-rift accommodation space was filled by amalgamated fluvial and lacustrine facies (Crayfish Subgroup) (Krassay *et al.*, 2004). A thick mudstone-rich volcaniclastic succession (Eumeralla Formation) was deposited during a decrease in tectonic activity during the Aptian and Albian (Krassay *et al.*, 2004). The main source rocks in the Otway Basin are of early Cretaceous age (Casterton Formation and Crayfish Subgroup in SA and Eumeralla Formation in Victoria), and the Pretty Hill Formation (Crayfish Subgroup) is the major reservoir unit in the SA Otway Basin (Bernecker *et al.*, 2003).

This basin and other parts of the southern Australian margin experienced a pulse of ~NW-SE directed shortening, uplift and exhumation during the late Albian-Cenomanian (Duddy, 1997, 2003; Hill *et al.*, 1995; MacDonald *et al.*, 2013), resulting in a regional mid-Cretaceous unconformity (MCU) (Fig. 3, 4). Major uplift in the eastern Otway Basin in the present-day Otway Ranges effectively isolated the Torquay sub-basin (Krassay *et al.*, 2004), whose late Cretaceous-Cenozoic stratigraphy shares a closer affinity with the Bass Basin to the east (Messent *et al.*, 1999). Western and central parts of the Otway Basin experienced renewed

~N-S to ~NE-SW directed rifting in the late Cretaceous (Perineck *et al.*, 1994), with some workers arguing for a significant oblique (sinistral) component (Schneider *et al.*, 2004). The locus of extension shifted to the south during the late Cretaceous, resulting in a series of ~NW-SE depocentres (the Morum and Nelson sub-basins) located to the south of the Tartwaup-Mussel fault zone (Bernecker *et al.*, 2003). Oblique-slip reactivation of N-S basement trends during ~NE-SW extension resulted in a series of ~N-S axial anticlinal and synclinal folds with a wavelength of ~5-15 km (Geary & Reid, 1998). The Upper Cretaceous Sherbrook Group generally comprises fluvial-deltaic and nearshore to shallow-marine siliclastic deposits, with the basal Waarre Formation acting as the major regional reservoir interval in the Victorian part of the Otway Basin.

Despite much research, the history of rifting and breakup between Australia and Antarctica remains poorly constrained (e.g. Williams *et al.*, 2011; Gibson *et al.*, 2013). It is generally agreed rifting in the Otway Basin culminated at the end of the Cretaceous and is marked by a regional intra-Maastrichtian unconformity (IMU), which is interpreted by some as marking the time of continental plate separation between Australia and Antarctica (Krassay *et al.*, 2004). However, there is no evidence for appreciable seafloor spreading off the Otway Basin at this time, with first ocean crust in the Otway Basin dated as middle Eocene (Norvick & Smith, 2001). Sedimentary successions in the Otway Basin become progressively more marine-influenced and calcareous throughout the Cenozoic, reflecting the progressive establishment of open marine circulation (McGowran *et al.*, 2004; Blevin & Cathro, 2008). Subsidence following the development of the intra-Maastrichtian unconformity initiated a major transgression across the Otway Basin resulting in deposition of the siliclastic Wangerrip Group (Bernecker *et al.*, 2003). Small growth wedges bounded by basinward-dipping reactivated late Cretaceous normal faults in the Portland Trough, onshore and offshore Victoria (Krassay *et al.*, 2004) imply that an extensional stress regime continued into the Palaeogene. The Wangerrip Group reaches a maximum thickness of >1200 m within the Portland Trough (Holdgate & Gallagher, 2003). It is separated by a major intra-Lutetian unconformity (ILU) associated with significant localised erosion from the overlying prograding nearshore to offshore marine clastics and carbonates of the late Eocene-early Oligocene Nirranda Group (Holdgate & Gallagher, 2003; Krassay *et al.*, 2004). This unconformity correlates with the onset of fast seafloor spreading in the southern Ocean at ~43 Ma (Veevers, 2000; McGowran *et al.*, 2004), and is associated with localised magmatic activity in both the Ceduna (Jackson, 2012) and Otway basins (Holford *et al.*, 2012). The Nirranda Group reaches its maximum thickness of ~200 m in two major depocentres, in the Port Campbell Embayment and Portland Trough (Holdgate & Gallagher, 2003). It is separated from the overlying Heytesbury Group by a regional intra-Oligocene unconformity (IOU) (Bernecker *et al.*, 2003). The late Oligocene-late Miocene Heytesbury Group (maximum thickness >1600 m) comprises marls and limestones deposited under fully marine conditions (Krassay *et al.*, 2004). A regional, late Miocene-Pliocene unconformity (MPU) (Dickinson *et al.*, 2002; Holford *et al.*, 2011*b*) separates the Heytesbury Group from the late Neogene succession of the Otway Basin which is characterised by relatively thin and localised mixed siliclastic-carbonate sediments and basaltic volcanic rocks that usually unconformably or disconformably overlie Heytesbury Group strata (Dickinson *et al.*, 2002; Tassone *et al.*, 2011).

Since the mid-Eocene onset of fast spreading ~43 Myr ago (Veevers, 2000; McGowran *et al.*, 2004), the Australian continent has been subjected to a largely compressional stress field resulting from the configuration of the Indo-Australian Plate boundaries (Fig. 1) (Sandiford & Quigley, 2009). This compression has caused widespread tectonic inversion in the sedimentary basins around and within the Australian continent (Etheridge *et al.*, 1991), and has had a profound influence on hydrocarbon occurrence by creating or amplifying traps in central Australia, the NW Shelf and the southern Australian margin (Hillis *et al.*, 2008*b*; Holford *et al.*, 2010*b*). With respect to the southern margin, compressional deformation has been documented most comprehensively in the Gippsland Basin, where several of Australia's largest hydrocarbon fields are located in inversion-related anticlines that mostly trend ~NE-SW or ~ENE-WSW (Hillis *et al.*, 2008*b*). Workers have identified key periods of deformation during the Eocene, Oligocene-early Miocene, and late Miocene-Pliocene (Brown, 1986; Dickinson *et al.*, 2001). A number of structures, including the anticline that hosts the Barracouta gas field, reveal evidence for multiple phases of growth (in this case, Oligocene-early Miocene and Plio-Pleistocene) (Dickinson *et al.*, 2001). From a seismic analysis of the late Oligocene-Recent Seaspray Group, Dickinson *et al.* (2001) identified two main structural styles that characterise Cenozoic compressional deformation in the Gippsland Basin: 1) reactivation and inversion of basin-margin normal faults with associated high amplitude hangingwall anticlines; and 2) broad low-relief anticlines in the main depocentre of the basin reflecting reactivation of deep fault blocks.

In comparison to the Gippsland Basin, the timing and distribution of Cenozoic compressional deformation in the Otway Basin is far less well understood. The most significant compressional structures occur onshore in the eastern Otway Basin, in and around the Otway Ranges, and in the adjacent Torquay sub-basin (Hill *et al.*, 1995; Dickinson *et al.*, 2002; Sandiford *et al.*, 2004; Clark *et al.*, 2012; Tassone *et al.*, 2012, 2013). Similar to the Gippsland Basin, these structures comprise a combination of ~NE-SW trending high and low-amplitude anticlines resulting from reactivation of syn-rift normal faults. Most studies have assigned a late Miocene-Pliocene age to the formation of these structures (Hill *et al.*, 1995; Dickinson *et al.*, 2002; Hillis *et al.*, 2008*b*) and other anticlines and reactivated faults indentified elsewhere in the Otway Basin (Perineck & Cockshell, 1995; Schneider *et al.*, 2004). In contrast to the Gippsland Basin, relatively few instances of Eocene-Oligocene growth of compressional structures in the Otway Basin have been reported. Exceptions include the Morum High on the outer continental shelf in the western Otway Basin, which Duddy *et al.* (2003) recognized as a mid-Eocene inversion structure.

SEISMIC DATABASE AND METHODOLOGY

With a few exceptions (e.g. Perineck & Cockshell, 1995), the majority of previous studies of structural styles in the Otway Basin have focused on individual structures (e.g. Schneider *et al.* 2004) or specifically on the South Australian (Jensen-Schmidt *et al.*, 2002) or Victorian (Hill *et al.*, 1995) sectors of the basin. With the aim of developing a better understanding of the basin-wide distribution and chronology of post-breakup compression, we have conducted seismic-stratigraphic and structural interpretation and mapping of a regional 2D seismic dataset that covers the length and breadth of the offshore Otway Basin (Fig. 2). The seismic data comprises part of PGS's Southern Australian Margin Digital Atlas (SAMDA), a compilation of ~113,000 line-kilometres of 2D seismic data from 129 separate surveys provided by federal (Geoscience Australia) and state governments (Primary Industries and Resources South Australia, and Department of Primary Industries Victoria). Our mapping of compressional structures in the Otway Basin utilizes ~30,000 line-kilometres of 2D seismic data from 14 individual surveys, and is calibrated to stratigraphic information from 17 offshore wells (Fig. 3).

In order to define the timing of post-breakup compression in the offshore Otway Basin we have conducted detailed mapping of three critical stratigraphic horizons that bound the major

Cenozoic supersequences of the Otway Basin (Fig. 5). These horizons we have mapped are; an intra-Maastrichtian (~67 Ma) unconformity (IMU) that occurs at the top of the Sherbrook Group and base of the Wangerrip Group; an intra-Lutetian (~46-43 Ma) unconformity (ILU) that occurs at the top of the Wangerrip Group and the base of the Nirranda Group; and, an intra-Oligocene (~29-28 Ma) unconformity (IOU) that occurs at the top of the Nirranda Group and the base of the Heytesbury Group (Fig. 3, 4; Boult *et al.*, 2002; Holdgate & Gallagher, 2003; Krassay *et al.*, 2004). Where possible, we have also mapped the distribution of the late Miocene-Pliocene (~10-5 Ma) unconformity (MPU) that marks the top of the Heytesbury Group (Dickinson *et al.*, 2002), though as this surface often occurs at shallow depths (less than several hundred metres) where the quality of available seismic data is often poor and biostratigraphic constraints are limited, it is not possible to confidently map the unconformity throughout the basin. A regional sequence-stratigraphic study by Krassay *et al.* (2004) which utilized biostratigraphic constraints from ~80 exploration wells showed that these unconformities can be traced throughout the onshore and offshore Otway Basin. All three unconformities are locally associated with severe erosion, and the intra-Maastrichtian and intra-Lutetian unconformities are at least in part tectonically controlled (Krassay *et al.*, 2004; McGowran *et al.*, 2004). The age-ranges of these unconformities are generally well constrained by planktonic foraminifera and palynological data from onshore and offshore exploration wells (McGowran *et al.*, 2004). These unconformities are broadly coeval with other unconformities and sequence boundaries of similar age that occur along almost the entire length of the southern Australian margin, from the Ceduna Basin in the west to the Gippsland Basin in the east (McGowran *et al.*, 2004; Blevin & Cathro, 2008).

POST-BREAKUP DEFORMATION STRUCTURES IN THE OTWAY BASIN

Here we describe results from mapping of seven anticlinal structures within the Cenozoic succession of the Otway Basin that formed during post-breakup compressional deformation. We applied conventional seismic-stratigraphic methodology (e.g. Mitchum *et al.*, 1977) for identifying and interpreting reflection terminations to map the intra-Maastrichtian, intra-Lutetian and intra-Oligocene unconformities. These dated horizons were then used to identify features such as thickness variations across hinge zones and onlapping reflections towards anticline limbs, to help constrain the age and chronology of individual structures related to compressional deformation. The amplitudes of the anticlines range from ~100 to 500 m and the lengths of their axial traces vary between ~4 and 30 km. Two dominant structural trends are identified, ~NE-SW and ~NW-SE.

Morum Anticline

The Morum Anticline is located ~65 km SW of Cape Jaffa, South Australia (Fig. 2b). This anticline is partially responsible for a structural culmination with significant bathymetric expression located near the continental shelf break, known as the Morum High. Local stratigraphic constraints are provided by the Morum-1 well, which intersected thin unconformity-bound Heytesbury (~138 m) and Wangerrip Group (~68 m) sequences overlying a thick Upper Cretaceous Sherbrook Group (~1656 m) succession (Duddy *et al.*, 2003). The well was drilled in 1975 and no significant shows were encountered.

The Morum Anticline trends ~NE-SW, has an axial trace ~19 km long and has a maximum amplitude of ~100 m (Fig. 6). Figure 6 shows that Cretaceous and older sediments in the fold are offset by a number of NW-trending normal faults, which form a series of fault blocks that run obliquely to the axis of the fold. Palaeocene-early Eocene Wangerrip Group sediments appear to thicken away from the crest of the fold and show no evidence for onlap onto the IMU. The top-Wangerrip Group unconformity (ILU) is an angular unconformity that separates tilted Wangerrip Group strata from relatively horizontal reflections in the overlying Heytesbury Group. A ~NNE-SSW-oriented seismic profile through the Morum Anticline shows clear truncations within the Wangerrip Group at the ILU to the north of the Morum-1 well (Fig. 7), where the Wangerrip Group is thicker. We interpret the Morum Anticline to have formed in mid-Eocene time, with the thinning of the Wangerrip Group section towards the crest of the structure indicating major intra-Lutetian erosion (cf. Duddy *et al.*, 2003). The absence of proven Nirranda Group sediments in the Morum-1 well, and the relatively thin Heytesbury Group succession implies that the Morum Anticline has acted as a structural high since the late Palaeogene.

Copa Anticline

The Copa Anticline is a structural culmination that comprises a series of small anticlines we have termed Copa A-C (Fig. 2b), located ~25 km off Beachport, South Australia. Copa A and B trend ~ENE-WSW and have axial traces ~6 km and ~2 km in length, respectively. Copa A has an amplitude of ~100 m (Fig. 8) whilst the amplitude of Copa C is ~60 m (measured at the IMU). These anticlines appear to be fault-propagation folds that formed following mild reverse-reactivation of underlying normal faults. The folds trend obliquely to the dominant ~NW-SE trending normal faults in the area (Fig. 2a, b), suggesting a component of dextral oblique-slip reactivation. The stratigraphy in this part of the basin is constrained by the Copa-1 well that was drilled in 1990 but detected no significant shows. It encountered thin Heytesbury Group (~304 m) and Wangerrip Group (~105 m) successions overlying a thick (~3,326 m) Sherbrook Group sequence (Fig. 3). The Wangerrip Group exhibits similar thicknesses across the limbs and crests of the anticlines, and there is little evidence for onlap onto the IMU indicating that these folds most likely formed subsequent to the early-Palaeogene. Heytesbury Group sediments appear to thin towards and onlap the ILU (Fig. 8), indicating mid-Eocene growth of Copa A and C.

The largest fold in the Copa structure is Copa B (Fig. 9). This anticline is located to the south of Copa A and has a ~NW-SE orientation. The axis of this anticline is oriented sub-parallel to a series of adjacent ~NW-SE trending normal faults, implying that the anticline might have formed following reverse-reactivation of these faults, although the faults still exhibit net-normal displacement (c.f. Williams *et al.,* 1989; Holford *et al.,* 2009*b*). There is some evidence for onlap of reflections near the base of the Wangerrip Group onto the IMU, but the largely consistent thickness of the Wangerrip Group across the fold implies that Copa C formed during the late Palaeogene-Neogene, rather than the late Cretaceous-early Palaeogene. Away from the crest of the anticline on its northeastern limb, late Eocene-Oligocene sediments are tilted and truncated beneath the IOU, implying that folding of Copa C occurred prior to the mid-Oligocene.

Pecten Anticline

The Pecten Anticline is a ~NE-SW trending fold that is ~18 km in length, and has an amplitude of ~270 m measured at the IMU. Several wells that have been drilled into this structure, targeting tilted syn-rift fault block traps, have discovered uncommercial (Pecten-1A) and commercial (Henry-1) gas accumulations (Fig. 2c). A ~NW-SE seismic profile through the Pecten Anticline reveals no evidence for onlap of Wangerrip Group, Nirranda Group and Heytesbury Group sediments onto the IMU, ILU and IOU, respectively. Furthermore, the thicknesses of the Wangerrip and Nirranda groups are largely consistent across the fold axis. These observations indicate that the Pecten Anticline formed subsequent to the Oligocene. Here the MPU is a localised angular unconformity marked by erosional truncation that separates underlying tilted strata from relatively horizontal overlying strata, suggesting that the folding event that formed the Pecten Anticline occurred in late Miocene-early Pliocene time. Figure 10 shows clear thinning of the section between the IOU and MPU towards the crest of the fold. The crest of the fold overlies the footwall of the underlying fault block, implying that the Pecten Anticline might have formed through a combination of differential compaction and shortening (cf. Gómez & Vergés, 2005). However, Figure 10 shows that the Pecten Anticline is one of several low-amplitude folds within the Cenozoic succession. These folds have a wavelength of ~20 km, and not all of the folds overlie Cretaceous footwall blocks. Thus whilst some of these folds may have been enhanced by differential compaction, we attribute their origin to layer-parallel shortening of the Cenozoic sedimentary succession

The Pecten Anticline appears to be broadly contiguous with the onshore Curdie Monocline (Fig. 2c), a probable late Miocene-Pliocene compressional structure that also trends ~NW-SE and has been mapped mainly from seismic sections (Tickell et al., 1992). This structure is indicated by a minor NW-dipping scarp, and appears to have formed above an east-dipping fault, which cross-sections (Fig. 11) indicate shows reverse-offset at top Otway, Sherbrook, Wangerrip and Nirranda levels (Edwards et al., 1996). Clark *et al.* (2011) suggest a tentative neotectonic slip rate of ~10–20 m Myr^{-1} for this fault based on displacement of late Miocene and younger stratigraphic markers.

Minerva Anticline

The Minerva Anticline is a domal structure located in the Shipwreck Trough (Fig. 2c), a ~N-S trending late Cretaceous depocentre where several significant gas discoveries have been made in the past few decades (Holford *et al.*, 2010*b*). Drilling of the Minerva-1 exploration well and the Minerva-2A appraisal well into the Minerva Anticline resulted in discovery of gas columns of 133 m and 111 m thickness, respectively, in Upper Cretaceous Sherbrook Group (Waarre Formation) sandstone reservoirs (Geary & Reid, 1998). Our mapping suggests that post-breakup growth of the Minerva Anticline is associated with the reverse-reactivation of a ~E-W striking, south-dipping normal fault. The growth history of this fold has previously been described by Schneider *et al.* (2004) who reported an initial, Campanian-Maastrichtian phase of growth, most probably related to transpressional deformation reported elsewhere in the Otway Basin at this time (Geary & Reid, 1998), and a subsequent Miocene-Recent phase of deformation (Schneider *et al.*, 2004). A ~NW-SE seismic profile through the anticline reveals no evidence for onlap onto the IMU or ILU, and similar to the Pecten Anticline, the Wangerrip and Nirranda groups exhibit little variation in thickness across the structure (Fig. 12). Folding of IOU and the observation of Heytesbury Group sediments onlapping the crest of the fold constrains the timing of post-breakup compressional deformation to beginning during the late Oligocene to early Miocene, consistent with the previous findings of Krassay *et al.* (2004) who suggested an early-mid Miocene timing for the onset of growth of the Minerva Anticline.

Similar to the offshore Pecten Anticline and the onshore Curdie Monocline, the Minerva Anticline may correlate with a major onshore fold, the ~NE-SW trending Ferguson Hill Anticline (Figs. 2c, 11) (Geary & Reid, 1998). However, the two folds appear to have rather different growth histories (Tickell *et al.*, 1992). Stratigraphic observations suggest that growth of the Ferguson Hill anticline began during the late Eocene-early Oligocene, based on the patchy distribution of Nirranda Group sediments in the Ferguson Hill area (Tickell *et al.*, 1992). A regional unconformity at the base of the Pliocene Hanson Plain Sand implies a subsequent phase of growth during the late Miocene-Pliocene (Tickell *et al.*, 1992). The Pliocene deposits have themselves been folded, and Sandiford (2003a) has used geomorphological evidence to suggest that the present-day topographic relief of the Ferguson Hill Anticline (where Pliocene strandlines are elevated at ~245 m above sea level) might largely be due to reverse-slip along ~NE-SW trending faults between ~2-1 Ma. Clark *et al.* (2011) estimated tentative neotectonic slip rates of ~23–58 m Myr^{-1} for the faults underlying this structure.

Point Ronald Anticline

The Point Ronald Anticline is a ~NE-SW trending anticline located ~10 km SE of the Minerva Anticline (Fig. 2c). The fold has an amplitude of ~330 m at the intra-Maastrichtian unconformity, and is ~6.5 km long. As with the Pecten and Minerva anticlines, seismic data through the fold show no clear evidence for onlap onto the IMU, ILU or IOU (Fig. 13). At deeper levels, thinning of sediment towards a mid-Cretaceous unconformity, suggests that a structural high related to compression may have existed at this location at this time. A number of workers have documented evidence for a major phase of compression, uplift and exhumation in the proximal Otway Ranges at ~95 Ma (Duddy, 1994; Hill *et al.*, 1995; Green *et al.* 2004). The MPU forms a seabed unconformity that truncates the underlying IOU. The thin sequence of post-late Oligocene sediments makes it difficult to constrain the age of the structure beyond Miocene or younger.

Crowes Anticline

This ~NE-SW trending anticline is a ~20 km long fold (Fig. 2c) that has maximum amplitude of ~430 m at the level of the MCU (Fig. 14). Onlap of Upper Cretaceous Sherbrook Group reflectors onto the MCU suggests that, similar to the Point Ronald Anticline, a structural high existed at this time, probably related to the mid-Cretaceous compression and uplift of the contiguous Otway Ranges (Duddy, 1994; Hill *et al.*, 1995). Indeed, this structure correlates with the onshore ~NE-SW trending Crowes Anticline (Fig. 2c), a major anticline within early Cretaceous Eumeralla Formation sediments in the SW Otway Ranges that has been mapped using both field observations and magnetic data (Fig. 11) (Edwards *et al.*, 1996). The IMU, ILU and IOU are all folded, implying a later phase of compression, although lack of onlap onto these unconformities indicates that none of them marks the timing of this compression (the IMU is marked by downlap of clinoforms representing the prograding delta systems of the Wangerrip Group). The MPU occurs near to the seabed, and is characterised by erosional truncation of underlying strata within the Heytesbury Group. We assign a late Miocene-early Pliocene age to the Cenozoic compression recorded by the Crowes Anticline. Folding most likely occurred in response to reverse-slip reactivation along a deep W-dipping normal fault imaged within the early Cretaceous succession (Fig. 14).

Loch Ard Anticline

The Loch Ard Anticline is a fold with a sinuous axial trace that broadly trends ~NE-SW, similar to other folds in the eastern Otway Basin (Fig. 2c). Fig. 14 shows a seismic profile through a ~NNW-SSE trending, ~6 km long segment of the structure that has an amplitude of ~100 m at the level of the IMU. Unlike the nearby Crowes Anticline, there is no evidence for onlap onto the deep mid-Cretaceous unconformity near the crest of this fold, implying that the Loch Ard Anticline was not growing at this time. However, there is evidence for onlap onto this unconformity on the outer limbs of the fold. In conjunction with the observation that the Upper Cretaceous Sherbrook Group succession reaches its maximum thickness along the axial plane, the Loch Ard Anticline appears to be a typical inversion structure, whereby a former depocentre has been converted into a structural culmination (e.g. Williams *et al.*, 1989). There is no onlap onto the ILU or IOU, although the observation that the Wanggerip Group succession encountered by the unsuccessful Loch Ard-1 well is much thinner than that penetrated by nearby wells (e.g. ~132 m compared with ~456 m at Minerva-1; Fig. 3) implies that the Loch Ard Anticline might have been a relative structural high throughout the Eocene. However, the main phase of growth clearly occurred during the late Miocene-early Pliocene, as evidenced by tilted and truncated reflections beneath the MPU.

THE NATURE, TIMING AND DURATION OF POST-BREAKUP COMPRESSION

Using a regional 2D seismic database we have constrained the Cenozoic growth history of a series of compressional folds in the offshore Otway Basin (Fig. 15). The Cenozoic folds have a variety of origins. Some folds, such as the Morum, Copa and Loch Ard anticlines have formed following reactivation of Cretaceous normal faults. In the east of the basin where the Cretaceous structural fabric broadly trends NE-SW to NNW-SSE (Fig. 2), folds such as the Minerva and Loch Ard anticlines may reflect the formation of growth anticlines within Cenozoic post-rift sediments due to reverse-slip reactivation of older normal faults. In the west of the basin, the Cretaceous structural fabric is more dominantly E-W to WNW-ESE and structures such as the Morum and Copa anticlines witness the oblique-slip reactivation of syn-rift normal faults with an east-west oriented, dextral shear sense (Perineck & Cockshell, 1995). However, most ‘inverted’ faults throughout the basin retain net-normal extension, with reverse-offset of syn or post-rift marker horizons rarely observed, and it is probable that most Cenozoic folds have been accentuated by the buttressing of syn-rift hangingwall sediments against rigid footwall blocks (cf. McClay, 1995). In addition, seismic data from the eastern Otway Basin reveal evidence for low-amplitude folding of the Cenozoic succession that is apparently unrelated to faulting (e.g. Fig. 10). Some of these folds may have been accentuated by differential compaction over pre-rift topography, but the regular spacing of these folds (wavelengths ~20 km) suggests that their origin is related to regional buckling of the Cenozoic sedimentary fill in response to relatively minor amounts of shortening. The increase in the amplitude of these folds from northwest to southeast within the basin is consistent with estimates of Cenozoic shortening from balanced regional cross-sections, which indicate negligible shortening in the western Otway Basin (e.g. Penola Trough), increasing to ~5% in the eastern Otway Basin (e.g. Otway Ranges) (Cooper & Hill, 1997). A basement control on Cenozoic deformation is most evident around the Otway Ranges, with the orientations of the reactivated Bambra and Torquay Faults that bound the Ranges to the NW and SE corresponding strongly to ~NE-SW striking structural trends in the underlying Selwyn Block (Cayley *et al.*, 2002).

A summary of the timing and duration of the growth of the folds in the Otway Basin is presented in Fig. 16. It is clear that post-breakup compression is not restricted to a single interval of time, with the main period of deformation in the western Otway Basin occurring during the mid-late Eocene (e.g. Morum and Copa), whilst in the eastern Otway Basin most structures were active during the late Miocene-Pliocene (e.g. Pecten, Crowes, Loch Ard). There is also evidence for Oligocene-Miocene folding in both western (e.g. Copa) and eastern parts of the basin (e.g. Minerva), whilst some structures appear to have experienced multiple periods of growth (e.g. Loch Ard). The onshore record of folding provides additional constraints on the history of post-breakup compression. The growth histories of some onshore folds appear to correlate both spatially and temporally with those we have described offshore (e.g. the Pecten Anticline and Curdie Monocline, the Crowes anticlines), and some folds may extend the temporal record of deformation within the basin (e.g. the Ferguson Hill Anticline which has evidence for growth during the late Eocene-early Oligocene and Pleistocene).

Defining the precise duration of the increments of fold growth is difficult based on available biostratigraphic constraints, but we suggest that the structures were generally active for a maximum of at least >1–10 Myr before they became quiescent and the locus of deformation migrated elsewhere within the basin. Palaeoseismic studies of faults with known Quaternary activity onshore Australia have identified episodic slip behaviour potentially analogous to that suggested for the Cenozoic structures discussed here, albeit over much shorter timescales, with brief episodes of activity separated by quiescent intervals of >10–100 Kyr (Crone *et al.*, 2003; Clark *et al.*, 2008; Leonard & Clark, 2011; Clark *et al.*, 2012). The approximate timescales of deformation we infer from the records of fold growth in the Otway Basin may represent an upper bound to the lifespan of intraplate compressional structures such as the 'active' faults documented onshore Australia, where surficial processes commonly limit the potential to constrain the extent of seismic cycles beyond the late Quaternary (<130 Ka) (Crone *et al.*, 2003).

CONSTRAINTS ON THE TIMING OF COMPRESSION FROM THERMOCHRONOLOGICAL DATA

Thermochronological data (e.g. apatite fission track analysis (AFTA), vitrinite reflectance (VR), apatite (U-Th)/He dating) acquired from exploration wells that targeted Cenozoic compressional structures provide further constraints on the timing of post-breakup compression. AFTA and VR data from Upper Cretaceous and Cenozoic units have led to identification of cooling episodes that correlate well with the timing of compression established from seismic mapping.

Duddy *et al.* (2003) analysed AFTA, VR and apatite (U-Th)/He data from the Morum-1 well (Fig. 17a) as part of a study into the thermal and structural history of the western Otway

Basin. This well drilled thin, unconformity-bound Heytesbury (~138 m) and Wangerrip Group (~68 m) sequences that overlie a thick Upper Cretaceous Sherbrook Group (~1656 m) succession (Duddy *et al.,* 2003). Palaeotemperatures determined from samples taken from the Upper Cretaceous succession are considerably higher than the present-day temperatures in the well as constrained by corrected bottom hole temperature (BHT) data (Duddy *et al.,* 2003). This indicates that the preserved section has been hotter in the past, and the form of the palaeogeothermal gradient defined by the data implies that deeper burial by ~1.5 km of now-eroded section, prior to cooling and exhumation, is the cause of the observed heating (Duddy *et al.,* 2003). Combined thermal history interpretation of several AFTA and apatite (U-Th)/He samples from the well suggests that the preserved Upper Cretaceous succession began to cool between 57 to 40 Ma (Duddy *et al.,* 2003). This latest Palaeocene-mid Eocene timing compares well with the age of the regional intra-Lutetian unconformity and is consistent with the thin package of Palaeocene-age Wangerrip Group sediments preserved beneath the Gambier Limestone (Duddy *et al.,* 2003). Unpublished AFTA and VR data from the Copa-1 well provide similar evidence for cooling and exhumation of Wangerrip Group (Dilwyn Formation) and older sediments beginning between 56 to 40 Ma (Geotrack International, 2003). Thus, thermal history data provide strong independent corroboration for the mid-Eocene deformation and indicate that the growth of the Morum (and likely Copa) anticlines was accompanied by substantial localised exhumation.

Unpublished thermal history data for two offshore wells located in the central Otway Basin (Breaksea Reef-1 and Bridgwater Bay-1; Fig. 2) was also available to this study. Both of these unsuccessful wells targeted faulted late Cretaceous-age anticlines, and Bridgwater Bay-1 was drilled on the southwestern edge of a large, southeasterly-plunging structural high that is visible on the intra-Maastrichtian time structure map (Fig. 5c). Although there is no clear evidence from seismic data for reactivation of these structures during the Cenozoic, thermal history data indicates that mid-Cenozoic cooling has occurred at both locations (Fig. 16). AFTA, VR and apatite (U-Th)/He data from Breaksea Reef-1 suggests that the Wangerrip Group (Dilwyn Formation) and older sequences began to cool from maximum palaeotemperatures ~10-20°C higher than present-day temperatures at some time in the time interval 55 to 30 Ma (Geotrack International, 2003). This cooling is attributed to uplift and erosion at the top-Wangerrip Group unconformity (ILU). AFTA and VR data from Bridgewater Bay-1 define a period of cooling from maximum palaeotemperatures 30°C higher than present-day temperatures beginning between 40 and 30 Ma (Geotrack International, 2003). This well contains a major unconformity that separates the mid-Eocene Nirranda Group from the overlying Miocene Port Campbell Limestone Formation (IOU). This unconformity is characterised on seismic data by truncation and downlap, and the AFTA and VR results are interpreted as recording significant erosion of the underlying Nirranda Group at the IOU (Geotrack International, 2003).

In the eastern Otway Basin, published palaeotemperature data from exploration wells and outcrops around the Otway Ranges are generally dominated by younger, late Cenozoic cooling (Duddy, 1997; Cooper & Hill, 1997; Green *et al.*, 2004; Holford *et al.*, 2011*b*), consistent with the predominance of younger (i.e. Miocene–Pliocene) fault reactivation and folding in this part of the basin. AFTA, VR and apatite (U-Th)/He data from the Nerita-1 well (Fig. 17b) in the contiguous Torquay sub-basin indicate that the development of the Nerita Anticline, a major ~NE-SW trending inversion structure commenced between 10 and 5 Ma (Fig. 18) (Holford *et al.*, 2011*b*), in excellent agreement with stratigraphic constraints provided by dating of the regional late Miocene-early Pliocene unconformity (Dickinson *et al.*, 2002). These data also indicate that the growth of the Nerita Anticline was accompanied by ~1 km of erosion (Fig. 17b) (Holford *et al.*, 2011*b*). AFTA, VR and apatite (U-Th)/He data from the nearby, onshore Anglesea-1 reveal similar evidence for significant exhumation (removing up to 750–950 m of Eocene-Miocene section) beginning in the late Miocene (between 12 to 7 Ma) (Green *et al.*, 2004). Both onshore and offshore in the Torquay sub-basin there is also evidence for older periods of post-breakup deformation, particularly during the mid-Eocene (Trupp *et al.*, 1994; Holdgate *et al.*, 2001).

Of the eastern Otway Basin compressional structures described in this study, the early-mid Miocene-age Minerva Anticline has the most comprehensive published thermal history dataset. Holford *et al.* (2010*b*) presented AFTA and VR data from Upper Cretaceous Sherbrook Group and Lower Cretaceous Otway Group samples recovered from the Minerva-1 exploration well (Fig. 17c). Palaeotemperatures estimated from these data are highly consistent with the present-day temperatures in the well constrained by BHT data, indicating that the preserved section is currently at or close to maximum post-depositional temperatures, and thus prohibiting independent timing constraints from thermal history data. In contrast to the Nerita and Morum anticlines, the growth of the Minerva Anticline accompanied by relatively minor coeval exhumation (<300 m), which may reflect deeper bathymetry during development of the latter structure (Holford *et al.*, 2010*b*). A study of AFTA and VR data from the La Bella-1 and Mussel-1 wells located on the Mussel Platform, approximately ~25 km to the south of the Minerva Anticline, indicated substantial localized exhumation beginning around 60 Ma, during which up to 1.5 km of section was removed (Duddy & Erout, 2001). This ‘pre-breakup’ exhumation is attributed to localized transpression during the opening of the Southern Ocean, which involved a significant component of left-lateral shear (Schneider *et al.*, 2004).

DRIVING MECHANISMS BEHIND POST-BREAKUP COMPRESSION IN THE OTWAY BASIN

Insights from contemporary stress data

The increasing recognition of post-breakup deformation along rifted continental margins has motivated numerous studies into the mechanisms that control such unanticipated tectonic activity, and a wide range of causal processes have consequently been invoked. These include the transmission of compressional stresses from active orogenic zones (Cobbold *et al.*, 2001; Ziegler *et al.*, 1995); topographic body forces such as ridge-push (Bott, 1993; Doré *et al.* 2008) or resulting from topographic gradients at continental margins (Pascal & Cloetingh, 2009); reactivation of basement lineaments (Lundin & Doré, 1996; Hudec & Jackson, 2002); and differential seafloor spreading and mantle drag (Mosar *et al.*, 2002; Le Breton *et al.*, 2012).

It is generally understood that the Quaternary-contemporary deformation of continental Australia and its rifted margins is controlled, to first-order, by a complex intraplate stress field generated by distant plate boundary interactions (Reynolds *et al.*, 2002; Sandiford, 2003b; Sandiford *et al.*, 2004; Hillis *et al.*, 2008*b*). Several key lines of evidence support this view. Present-day maximum horizontal stress orientations across Australia, measured using earthquake focal mechanisms, borehole breakouts and drilling-induced tensile fractures (DITFs), indicate consistent stress orientations within individual regions and basins, but considerable variation from region-to-region across the continent (Fig. 1) (Hillis & Reynolds, 2000). This complex pattern of intraplate stress orientations contrasts strongly with comparable continental regions (e.g. western Europe; Gölke & Coblentz, 1996), which tend to show maximum horizontal stress orientations that broadly parallel the direction of plate motion. Independent studies of Quaternary-contemporary faulting across Australia indicate that the strikes of many of the identified structures are broadly orthogonal to the present-day stress orientations determined in specific regions (Sandiford, 2003b; Quigley *et al.*, 2006; Hillis *et al.*, 2008*b*). It is notable that, where reliable kinematic constraints are available, onshore Quaternary faults generally indicate dip-slip or oblique-slip reverse motions, and no faults with demonstrable motions that are purely strike-slip or normal have been documented as yet (Hillis *et al.,* 2008*b*).

Despite the evidently complicated pattern of contemporary stress orientations across Australia, valuable insights into the origins of this stress field have been obtained from two-dimensional, elastic finite element modelling applied to the Indo-Australian Plate (Coblentz *et al.,* 1998; Reynolds *et al.,* 2002; Sandiford *et al.,* 2004). These models have achieved excellent statistical fits between observed and predicted stress orientations by focussing the distributed ‘ridge-push’ plate-driving torques related to the SW boundary of the Indo-Australian Plate against the collisional segments of the complex convergent NE plate boundary that act as sources of resistance (i.e. Himalayas, New Guinea, New Zealand) (Hillis *et al.*, 2008*b*; Sandiford & Quigley, 2009). The excellent agreement between observed present-day stress patterns and those predicted from modelling studies lends strong credence to the notion that plate boundary forces exert the primary control upon the contemporary stress field, and thus deformation, of Australia and its margins. An important corollary of this premise is that, although individual plate boundaries and their associated driving or resisting torques unquestionably exert a profound influence on the state of stress within the plate interior, in the case of Australia at least, both the contemporary stress and deformation fields cannot be attributed to a single plate boundary force or event.

Present-day maximum horizontal stress orientations in the Otway Basin have been well-constrained using borehole breakouts and drilling induced tensile fractures (DITFs) interpreted from petroleum wells (Hillis *et al.,* 1995; Nelson *et al.,* 2006). Maximum horizontal stress orientations are generally oriented NW-SE, with data revealing a slight anticlockwise rotation from ~125°N in the western (SA) Otway Basin, to ~135°N in the east (Victoria) (Fig. 15) (Nelson *et al.*, 2006). Farther to the east, in the Gippsland Basin, there is additional rotation of the maximum horizontal stress orientation to ~139 °N (Nelson *et al.,* 2006). The broad ~NW-SE strike directions of the majority of the anticlines documented in this study are thus consistent with independently determined present-day stress orientations, implying that post-breakup deformation in the Otway Basin has been controlled by plate-boundary forces.

There is less certainty regarding stress magnitudes and the nature of the stress contemporary regime in the Otway Basin. The neotectonic record, particular from the eastern Otway Basin, favours a reverse-fault stress regime (i.e. $\sigma_H > \sigma_h > \sigma_v$) (Sandiford *et al.*, 2004; Hillis *et al.,* 2008*b*), but petroleum data from wells located throughout the basin suggest a strike-slip fault stress regime (i.e. $\sigma_H > \sigma_v > \sigma_h$) (Nelson *et al.,* 2006). Using density logs to calculate vertical stress magnitudes, leak-off tests to constrain minimum horizontal stress and DITFs to calculate maximum horizontal stress magnitudes, Nelson *et al.* (2006) found the state-of-stress to be strike-slip in both the SA and Victorian parts of the Otway Basin. Their analysis revealed increases in both minimum and maximum horizontal stress magnitudes from western to eastern parts of the basin (minimum horizontal stress increasing from ~15.5 MPa km^{-1} to ~18.5 MPa km^{-1}; maximum horizontal stress increasing from ~29 MPa km^{-1} to ~37 MPa km^{-1}). The high magnitude of minimum horizontal stress observed in the eastern Otway

Basin suggests that the state-of-stress is approaching a reverse fault stress regime (Nelson *et al.,* 2006).

Several previous studies have interpreted the record of Cenozoic deformation along the eastern parts of the southern Australian margin in terms of an E-W orientated dextral strike-slip regime that resolves into ~NW-SE directed compression (Perineck & Cockshell, 1995; Jensen-Schmidt *et al.,* 2002; Pollock, 2003), consistent with present-day maximum horizontal stress orientations in the Otway Basin inferred from petroleum data (Hillis *et al.,* 1995; Nelson *et al.,* 2006). A dextral strike-slip regime can account for the formation of ~NW-SE oriented anticlines that form under compression, or through transpressional shortening during the reactivation of ~WNW-ESE oriented normal faults, such as those which dominate the structural grain in the western Otway Basin (e.g. around the Morum and Copa anticlines).

However, a recent reanalysis of stress magnitude data from the Otway Basin (King *et al.,* 2012) using a new method for evaluating leak-off test results that assumes shear failure of pre-existing fractures under compressional stress regimes (Couzens-Schultz & Chan, 2010) suggests that the present-day state-of-stress in the Otway Basin is better characterized by a reverse-fault stress regime than the strike-slip fault stress regime proposed in earlier studies (e.g. Hillis *et al.,* 1995; Nelson *et al.,* 2006). Such a stress regime would appear to be more compatible with the neotectonic faulting record from southern Australia, and with independent constraints on the state-of-stress from earthquake focal mechanism solutions (Hillis *et al.,* 2008*b*).

Linking post-breakup compression to evolving plate boundary forces

The recognition of spatially and temporally migrating post-breakup deformation in the Otway Basin is one of the key outcomes of this study. Figure 16 depicts a chronology of the timing and duration of activity for the structures we have identified. This confirms that deformation within the basin has been common following breakup, but is mainly concentrated into several periods of enhanced deformation when growth is evident on multiple structures within the basin. Here we discuss the significance of these enhanced periods of deformation, which occurred during the mid-Eocene and the late Miocene-early Pliocene. During the latter there is evidence for active growth of the Pecten, Ferguson Hill, Point Ronald, Crowes and Loch Ard anticlines in the Otway Basin (Fig. 16), the Nerita Anticline in the adjacent Torquay sub-basin (Holford *et al.*, 2010*b*, 2011*b*), and multiple onshore neotectonic faults that displace late Miocene-Quaternary sedimentary and volcanic rocks (Clark *et al.,* 2012). The growth of these structures is coeval with the development of a widespread low-angle unconformity that is dated as between 10 and 5 Myr old, and occurs over a distance of ~1000 km along the southern Australian margin, from the Gulf St Vincent Basin in the NW to the Gippsland Basin in the SE (Dickinson *et al.*, 2001, 2002; Sandiford *et al.*, 2004).

In an investigation into the causes of neotectonic activity in southeastern Australia, Sandiford *et al.* (2004) concluded that the origin of the present-day stress field in this part of the Australian continent can be traced back to the terminal Miocene or early Pliocene (~10–5 Ma). They argued that a significant change in the state-of-stress occurred at this time in response to concomitant changes in the coupling of the Indo-Australian and Pacific plates following the building of the Southern Alps in New Zealand, from ~12 Ma onwards (Sandiford *et al.,* 2004). They illustrated the considerable influence of the New Zealand plate boundary segment on the southeastern Australian stress regime through the use of finite element models similar to those constructed by Reynolds *et al.* (2002). Their modelling results indicate that in the absence of any transmission of force from the New Zealand boundary to the plate, maximum horizontal stress orientations in the Otway and Gippsland basins would most likely have ~NE-SW-trending azimuths. With increasing magnitudes of collisional force, modelled maximum horizontal stresses progressively rotate to ~NW-SE orientations that are consistent with both present-day stress data and the strikes of the post-breakup deformation structures documented herein. These results also suggest that the observed anticlockwise rotation of maximum horizontal stress that occurs between the western Otway (~125°N) and Gippsland basins (~139°N) can be understood in terms of relative proximity to the New Zealand boundary, which may also control the increase in minimum and (inferred) maximum horizontal stress magnitudes towards eastern parts of the southern margin (Sandiford *et al.*, 2004; Nelson *et al.*, 2006).

It is important to acknowledge that important changes to the nature of the Indo-Australian plate boundary zones in late Miocene-early Pliocene time were not restricted to the New Zealand region. As summarised by Hillis *et al.* (2008*b*), this period witnessed the collision between the Ontong Java Plateau and the Solomon Arc (Wessel & Kroenke, 2000), the initiation of compressional deformation in the Indian Ocean (Krishna *et al.*, 2001) and the onset of transpressional deformation and uplift in New Guinea (Hill & Hall, 2003). The latter deformation event, which occurred some 3500 km to the north of this study area, has previously been invoked as the key incident responsible for post-breakup deformation in the Otway Basin (Hill *et al.*, 1995).

Our preferred explanation for the late Miocene-early Pliocene deformation pulse, based on the insights obtained from finite element modelling of the contemporary Australian stress field (e.g. Reynolds *et al.*, 2002; Müller *et al.*, 2012), is that it represents an intraplate response to the reconfigurations of the Indo-Australian Plate boundaries which all combine to influence the balance of forces within the plate. However, the recognition of Eocene to early-Miocene age post-breakup structures striking ~NE-SW shows that the ~NW-SE σ_H orientation is a more persistent feature than proposed by Sandiford *et al.* (2004), who suggested that this stress orientation was initiated at ~12 Ma by increased coupling of the Australian-Pacific plate boundary. This observation implies that factors other than the Miocene evolution of the Australian-Pliocene plate boundary must be responsible for the ~NW-SE σ_H orientation, though it remains possible that the late Miocene-early Pliocene deformation pulse reflects a significant increase in stress magnitudes resulting from the enhanced forcing from New Zealand (Sandiford *et al.*, 2004).

The second period of enhanced post-breakup deformation that merits discussion occurred during the mid-Eocene. In the western Otway Basin, this time interval witnessed significant growth on the Morum and Copa anticlines (Figs. 15, 16) as indicated by mapping of a regionally extensive intra-Lutetian unconformity, and supported by thermochronological data which record substantive localised exhumation in the vicinity of these structures (Duddy *et al.*, 2003). In the eastern Otway Basin there is evidence that the Loch Ard Anticline underwent uplift at this time based on observed thinning of the Wangerrip Group across this structure. There is also evidence for mid-Eocene fault reactivation in the Port Campbell Embayment from structural mapping of onshore 3D seismic data (P. Boult personal communication), and folding associated with the intra-Lutetian unconformity has been shown to occur both offshore and onshore in the adjacent Torquay sub-basin (Trupp *et al.*, 1994; Holdgate *et al.*, 2001). Given the pervasive extent of mid-Eocene deformation, it may be the case that neotectonic faults for which only late Miocene-onwards displacements can be confidently demonstrated (Clark *et al.*, 2012), may have also accrued slip during the mid-Eocene (and/or Oligocene-early Miocene).

The Morum Anticline strikes ~NE-SW, consistent with the orientations of structures active during the late Miocene-early Pliocene (and those active during the Oligocene-early Miocene; Fig. 15), implying that maximum horizontal stress in the basin has broadly trended ~NW-SE for the past ~45 Ma. The folds that comprise the Copa Structure have varying trends including ~NNE-SSW and ~NW-SE, but the structure is generally consistent with having formed by oblique-slip reactivation of normal faults under ~NW-SE directed compression.

Our best estimate for the timing of enhanced deformation during the mid-Eocene based on stratigraphic constraints and mapping of individual structures is ~45 to 38 Ma. In a similar fashion to the subsequent phase of basin-wide deformation during the late Miocene-early Pliocene, fault-reactivation and folding during the mid-Eocene was also contemporaneous with major plate boundary reconfigurations. Most relevant in terms of the evolution of the southern margin, final plate separation between Australia and Antarctica is thought to have been achieved by ~43 Ma (Norvick & Smith, 2001). Breakup marked the onset of fast spreading in the Southern Ocean, with half-spreading rates along the Great Australian Bight ridge system accelerating from <0.5 cm yr^{-1} prior to 43 Ma, to 3.3 cm yr^{-1} by 34 Ma (Li *et al.*, 2003). Several other significant plate boundary adjustments took place at approximately 43 Ma, including an acceleration and ~70° counter clockwise rotation of motion of the Pacific Plate, and the abandonment of the eastern Indian Ocean spreading ridge leading to the fusing of the Indian and Australian plates (Veevers, 2000). These events, which followed the cessation of spreading in the Tasman Sea at ~52 Ma (Gaina *et al.*, 1998) and accompanied the deceleration of the Indian Plate during the nascent stages of Himalayan collision (Patriat & Achache, 1984), conspired to establish a compressional stress regime within the Australian crust dominated by ridge push to the south of Australia and increased resisting forces to the north (Dyksterhuis & Müller, 2008; Sandiford & Quigley, 2009; Müller *et al.,* 2012). We submit that the mid-Eocene shortening in the Otway Basin and adjacent parts of the southern margin represents an intraplate response to the initiation of a compressional stress regime within the Indo-Australian Plate shortly after breakup.

COMPARISONS WITH THE RECORD OF POST-BREAKUP DEFORMATION AT THE NW EUROPEAN ATLANTIC RIFTED MARGIN

We conclude this discussion with a brief comparison between the records of post-breakup deformation at the southern Australian margin and NW European Atlantic margin, which contains the best-documented record of post breakup deformation at a rifted continental margin. The NW European margin formed following multiple Permian to Paleogene rifting episodes, with continental breakup between NW Europe and Greenland achieved by ~53.7 Ma (Doré *et al.*, 2008). The post-breakup sedimentary record of the margin is dominated by multiple, Eocene and younger, siliclastic sediment wedges that prograde from the continental shelves of the British Isles, Norway and Faroe Islands, accompanied by the deposition of deep-water contourites in the adjacent basins since late Eocene time (Stoker *et al.,* 2010). Previous studies have documented widespread compressional folding on the Rockall-Faroes-West Shetland and mid-Norwegian sections of this margin, with particularly intense phases of deformation identified during the mid-Eocene to Oligocene and early-mid Miocene (Boldreel and Anderson, 1998; Lundin and Doré, 2002; Stoker *et al.,* 2005).

Major post-breakup structures on the mid-Norwegian margin are generally located within the Vøring Basin, with compressional features largely absent from the adjacent Møre Basin and Halten Terrace (Doré *et al.*, 2008) The structures comprise large-scale, elongate anticlines with four-way dip closure and low amplitude/wavelength ratios that mostly trend ~NE-SW to N-S (Gómez & Vergés, 2005). The largest structures include the Ormen Lange Dome and Helland-Hansen Arch, the latter of which has an amplitude of ~1 km and axial trace length of ~200 km (Doré *et al.*, 2008). Both these structures have documented mid-Eocene-early Oligocene growth phases and were reactivated during the early-mid Miocene when other folds including the Vema, Hedda, and Naglfar domes also developed (Lundin and Doré, 2002). Though some structures such as the Helland Hansen Arch appear to be related to the reactivation of normal fault systems, many of the structures do not appear to be directly linked to underlying faults (Doré *et al.*, 2008). Whilst a compressional origin of these structures is generally agreed (Doré *et al.*, 2008), shortening magnitudes appear to be minor (a few percent) (Vågnes *et al.*, 1998) and in many cases the amplitudes of folds have been enhanced by sedimentation and differential compaction on their flanks (Gómez & Vergés, 2005).

Post-breakup structures in the Rockall-Faroes-West Shetland area also mostly comprise folds that exhibit a wide variation in scale (axial trace lengths <10 to >250 km), orientation and timing of growth (Ritchie *et al.*, 2008). The largest structures include the Fugloy, Munkagrunnur and Wyville Thomson ridges, which form prominent bathymetric highs with reliefs of ≤900 m (Stoker *et al.*, 2005). Seismic mapping reveals significant early-mid Miocene growth of these structures (Stoker *et al.*, 2005), but also reveals long-lived growth of the latter throughout the Eocene-Oligocene (Ritchie *et al.*, 2008). Other structures with long-lived growth histories include the North Hatton Bank Fold Complex on the Hatton High (mid-Eocene to early Oligocene). In the NE Faeroe-Shetland Basin there are numerous NE-SW and NNE-SSE trending growth folds that developed in the early-mid Miocene and early Pliocene, and some of the younger structures have associated raised seabed profiles suggesting ongoing compression (Ritchie *et al.*, 2008).

In comparison to the NW European Atlantic margin, both the quantity and dimensions of post-breakup structures along the southern Australian margin are smaller (Tuitt *et al.*, 2011). However, at both margins the orientations of paleostresses inferred from compressional structures are highly consistent with independent determinations of the present-day stress field. Post-breakup structures on the NW European margin exhibit larger variation in orientation, which at least partially reflects the protracted and complex rifting history of this margin and the influence of syn-breakup magmatism (Ritchie *et al.*, 2008). Fold trends show most consistency in the Faeroe-Shetland Basin (mostly ~NE-SW) and offshore Norway (broadly ~N-S), and paleostress orientations from these structures are also largely consistent with the observed ~NW-SE present-day maximum horizontal stress orientation (Gölke and Coblentz, 1996). Furthermore, both margins display similarities in the chronology of deformation, with clear periods of enhanced compressional activity (e.g. early-mid Miocene in NW Europe, late Miocene-early Pliocene in southern Australia), but fold growth has taken place over extended periods at both margins following breakup. Doré *et al.* (2008) suggest that the mid-Miocene acme of deformation along the NW European margin can be explained by enhanced body forces, and thus intraplate stress magnitudes, resulting from development of the Iceland insular margin on the North Atlantic ridge-system during the middle Miocene. This enhanced ridge-push forcing, combined with the presence of hyperextended and weakened lithosphere (Lundin & Doré, 2011), may account for the larger dimensions of post-breakup structures on the NW European margin.

CONCLUSIONS

Seismic mapping in the offshore Otway Basin has identified a number of post-breakup compressional structures in this supposedly 'passive' rifted margin setting. Individual structures have a variety of ages but taken as a whole, the evidence presented here shows that the Otway Basin has been subject to continual compressional forcing subsequent to breakup, with strain migrating spatially and temporally within the basin and often reactivating syn-breakup fault systems. However, most post-breakup compressional activity is concentrated into two periods of enhanced deformation during the mid-Eocene and late Miocene-Pliocene. Our findings have important implications for the understanding of 'passive' margin tectonism and for the evolution of intraplate stress fields. Unanticipated deformation of 'passive' margins has been increasingly documented as petroleum exploration progressively focuses on deep-water basins along continental shelves worldwide (e.g. Cobbold *et al.*, 2001; Hudec & Jackson, 2002; Doré *et al.*, 2008). Attempts to explain such deformation have frequently focused on the episodic nature of post-breakup fault-reactivation, folding and uplift, and have attempted to link periods of tectonic activity with far-field events at plate boundaries (e.g. Ziegler *et al.*, 1995; Holford *et al.*, 2009*a*).

The chronology of post-breakup deformation that we have established does indeed reveal several periods of enhanced fault-reactivation and folding, occurring immediately after mid-Eocene breakup and subsequently during the late Miocene-early Pliocene, though growth on some important structures has clearly occurred outside of these time intervals, suggesting that post-breakup compression may be more common than is generally assumed. The key periods of enhanced post-breakup deformation clearly correlate with significant events at proximal Indo-Australian Plate boundaries (e.g. final Australia-Antarctica plate separation and onset of fast spreading in the mid-Eocene, development of the Southern Alps along the New Zealand plate boundary segment), and these events should be viewed in terms of larger-scale reconfigurations of the Indo-Australian Plate boundaries that occurred at these times. These reconfigurations served to establish a compressional stress regime within the plate interior, with contemporary Australia characterised by unusually high levels of seismicity for a continental region far from plate boundaries, and demonstrably high horizontal stress magnitudes in the upper crust (Hillis *et al.*, 2008*b*; Sandiford & Quigley, 2009).

We believe that the record of post-breakup deformation in the Otway Basin can be understood best in terms of the responses of the passive margin sedimentary succession and rift-related fault-systems to a continuously evolving compressional stress field that is coupled with, and controlled by, the progressive evolution of plate boundaries and their associated driving or resisting forces. Periods of enhanced basin-wide deformation are likely related to step changes in the intraplate stress regime (i.e. magnitudes and/or orientations) resulting from significant modifications in the nature or disposition of one or more plate boundary segments. This theory appears to be consistent with the record of post-breakup deformation along the NW European Atlantic passive margin, which contains similar evidence for protracted fault-reactivation and folding since ~53 Ma, but with periods of widespread compression in discrete time intervals that coincide with fundamental changes in plate boundary forcing (Doré *et al.*, 2008; Ritchie *et al.*, 2008; Holford *et al.*, 2009*a*). To obtain further insights into the nature and origin of post-breakup deformation at rifted margins, it will be necessary to compare the records of the southern Australian and NW European margins with deformation chronologies obtained at other passive margins that appear to be uncharacteristically active, such as the conjugate margins of the South Atlantic Ocean (e.g. Cobbold *et al.*, 2001; Hudec & Jackson, 2002; Turner *et al.*, 2008).

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REFERENCES

BERNECKER, T., SMITH, M.A., HILL, K.A. & CONSTANTINE, A.E. (2003) Oil and gas, fuelling Victoria's economy. In: *Geology of Victoria* (Ed. by W.D. Birch), *Geol. Soc. Aust., Spec. Publ.*, 23, 469-487.

BLEVIN, J. & CATHRO, D. (2008) Australian Southern Margin Synthesis. *Client report to Geoscience Australia by FrOG Tech Pty Ltd.* Project GA707,

BOLDREEL, L.O. & ANDERSEN, M.S. (1998) Tertiary compressional structures on the Faroe-Rockall Plateau in relation to northeast Atlantic ridge-push and Alpine foreland stresses. *Tectonophys.* 300, 13-28.

BOND, G.C. & KOMINZ, M.A. (1988) Evolution of thought on passive continental margins from the origin of geosynclinal theory (~1860) to the present. *Geol. Soc. Am., Bull.*, 100, 1903-1933.

BOTT, M.H.P. (1993) Modelling the plate-driving mechanism. *J. Geol. Soc.*, 150, 941-951.

BOULT, P.J., WHITE, M.R., POLLOCK, R., MORTON, J.G.G., ALEXANDER, E.M. & HILL, A.J. (2002) Chapter 6: Lithostratigraphy and environments of deposition. In: *Petroleum Geology of South Australia. Vol. 1: Otway Basin, Second Edition.* (Ed. by P.J. Boult and J.E. Hibburt).

BRAUN, J. (2010) The many surface expressions of mantle dynamics. *Nat. Geosci.*, 3, 825-833.

BROWN, B.R. (1986) Offshore Gippsland Silver Jubilee. In: *Second South-Eastern Australia Oil Exploration Symposium, Melbourne* (Ed. by R.C.Glenie), pp. 29–56. The Petroleum Exploration Society of Australia Victoria and Tasmanian Branch, Canberra, Australia.

CAYLEY, R.A., TAYLOR, D.H., VANDENBERG, A.H.M. & MOORE. D.H. (2002) Proterozoic-early Palaeozoic rocks and the Tyennan Orogeny in central Victoria: the Selwyn Block and its tectonic implications. *Aust. J. Earth Sci.*, 49, 225-254.

CÉLÉRIER, J., SANDIFORD, M., HANSEN, D.L. & QUIGLEY, M. (2005) Modes of active intraplate deformation, Flinders ranges, Australia. *Tectonics*, 24, 10.1029/2004TC001679.

CLARK D., DENTITH M., WYRWOLL K.H., YANCHOU L., DENT V. & FEATHERSTONE, C. (2008) The Hyden fault scarp, Western Australia: paleoseismic evidence for repeated Quaternary displacement in an intracratonic setting. *Aust. J. Earth Sci.*, 55, 379-395.

CLARK, D., McPHERSON, A. & COLLINS, C.D.N. (2011) *Australia's seismogenic neotectonic record: a case for heterogeneous intraplate deformation.* Record 2011/11, Geoscience Australia, Canberra.

CLARK, D., McPHERSON, A. &VAN DISSEN, R. (2012) Long-term behaviour of Australian stable continental region (SCR) faults. *Tectonophysics,* 566-567, 1-30.

COBBOLD, P.R., MEISLING, K. & MOUNT, V. S. (2001) Reactivation of an obliquely rifted margin, Campos and Santos basins, southeastern Brazil. *AAPG Bull. (Am. Assoc. Petrol. Geol.),* 85, 1925-1944.

COBBOLD, P.R., ROSSELLO, E.A., ROPERCH, P., ARRIAGADA, C., GÓMEZ, L.A. & LIMA, C. (2007) Distribution, timing, and causes of Andean deformation across South America. In: *Deformation of the Continental Crust: The Legacy of Mike Coward* (Ed. by A.C. Ries, R.W.H. Butler and R.H. Graham), *Geol. Soc. Spec. Publ.*, 272, 321-343.

COBBOLD, P.R., GILCHRIST, G., SCOTCHMAN, I., CHIOSSI, D., FONSECA CHAVES, F., GOMES DE SOUZA, F. & LILLETVEIT, R. (2010) Large submarine slides on a steep continental margin (Camamu Basin, NE Brazil). *J. Geol. Soc.*, 167, 583-592.

COBLENTZ, D.D., ZHOU, S., HILLIS, R.R., RICHARDSON, R.M. & SANDIFORD, M. (1998) Topography, boundary forces, and the Indo-Australian intraplate stress field. *J. Geophys. Res.*, 103, 919–931.

COOPER, G.T. & HILL K.C. (1997) Cross-section balancing and thermochronological analysis of the Mesozoic development of the eastern Otway Basin. *APPEA J.,* 37, 390-414.

COUZENS-SCHULTZ, B.A. & CHAN, A.W. (2010) Stress determination in active thrust belts: An alternative leak-off pressure interpretation. *J. Struct. Geol.,* 32, 1061-1069.

CRONE, A.J., DE MARTINI, P.M., MACHETTE, M.N., OKUMURA, K. & PRESCOTT, J.R. (2003) Paleoseismicity of two historically quiescent faults in Australia: Implications for fault behaviour in stable continental regions. *Bull. Seis. Soc. Am.,* 93, 1913-1934.

DAVIES, R., CLOKE, I., CARTWRIGHT, J., ROBINSON, A. & FERRERO, C. (2004) Post-breakup compression of a passive margin and its impact on hydrocarbon prospectivity: An example from the Tertiary of the Faeroe–Shetland Basin, United Kingdom. *AAPG Bull. (Am. Assoc. Petrol. Geol.),* 88, 1-20.

DENHAM, D., WEEKES, J. & KRAYSHEK, C. (1981) Earthquake evidence for compressive stress in the southeast Australian crust. *Aust. J. Earth Sci.,* 28, 323-332.

DICKINSON, J.A., WALLACE, M.W., HOLDGATE, G.R., DANIELS, J., GALLAGHER, S.J. & THOMAS, L. (2001) Neogene tectonics in SE Australia: implications for petroleum systems. *APPEA J.*, 41, 37-52.

DICKINSON, J.A., WALLACE, M.W., HOLDGATE, G.R., GALLAGHER, S.J. & THOMAS, L. (2002) Origin and timing of the Miocene-Pliocene unconformity in southeast Australia. *J. Sed. Research,* 72, 288–303.

DORÉ, A.G., CORCORAN, D.V. & SCOTCHMAN, I.C. (2002) Prediction of the hydrocarbon system in exhumed basins, and application to the NW European margin. In: *Exhumation of the North Atlantic Margin: Timing, Mechanisms and Implications for Petroleum Exploration* (Ed. by A.G. Doré, J.A. Cartwright, M.S. Stoker, J.P. Turner & N. White), *Geol. Soc. Spec. Publ.,*196, 401-429.

DORÉ, A.G., LUNDIN, E.R., KUSZNIR, N.J. & PASCAL, C. (2008) Potential mechanisms for the genesis of Cenozoic domal structures on the NE Atlantic margin: pros, cons, and some new ideas. In: *The Nature and Origin of Compression in Passive Margins* (Ed. by H. Johnson, A.G. Doré, R.W. Gatliff, R. Holdsworth, E.R. Lundin, and J.D. Ritchie), *Geol. Soc. Spec. Publ.*, 306, 1-26.

DUDDY, I.R. (1994) The Otway Basin: thermal, structural and tectonic and hydrocarbon generation histories. In: *NGMA/PESA Otway Basin Symp.*,(Ed. by D.M. Finlayson), AGSO record 1994/14, 35–42.

DUDDY, I.R. (1997) Focussing exploration in the Otway Basin: understanding timing of source rock maturation. *APPEA J.,* 37, 178-191.

DUDDY, I.R. (2003) Mesozoic, a time of change in tectonic regime. In: *Geology of Victoria* (Ed. by W.D. Birch), *Geol. Soc. Aust., Spec. Publ.,* 23, 239-286.

DUDDY, I.R. & EROUT, B. (2001) AFTA-calibrated 2-D modelling of hydrocarbon generation and migration using Temispack: preliminary results from the Otway Basin. In: *Eastern Australian Basins Symposium 2001* (Ed. by K.C. Hill & T. Bernecker), *Petrol. Soc. Aust., Spec. Publ.,* 403–412.

DUDDY, I.R, EROUT, B., GREEN, P.F., CROWHURST, P.V. & BOULT, P.J. (2003) Timing constraints on the structural history of the western Otway Basin and implications for hydrocarbon prospectivity around the Morum High, South Australia. *APPEA J.*, 43, 59-83.

DYKSTERHUIS, S. & MÜLLER, R.D. (2008) Cause and evolution of intraplate orogeny in Australia. *Geology*, 36, 495-498.

EDWARDS J., LEONARD J.G., PETTIFER G.R. & MCDONALD P.A. (1996) Colac 1 : 250 000 map geological report. *Geological Survey of Victoria. Report* 98.

ETHERIDGE, M., MCQUEEN, H. & LAMBECK, K. (1991) The role of intraplate stress in Tertiary (and Mesozoic) deformation of the Australian continent and its margins: a key factor in petroleum trap formation. *Explor. Geophys.*, 22, 123-128.

GAINA, C., MÜLLER, D.R., ROYER, J-Y., STOCK, J., HARDEBECK, J., & SYMONDS, P. (1998) The tectonic history of the Tasman Sea: A puzzle with 13 pieces. *J. Geophys. Res.*, 103, 12413-12433.

GEARY, G.C. & REID, I.S.A. (1998) Geology and prospectivity of the offshore eastern Otway Basin, Victoria. *Victorian Initiative for Minerals and Petroleum Report*, 55.

GIBSON, G.M., MORSE, M.P., IRELAND, T.R. & NAYAK, G.K. (2011) Arc-continent collision and orogenesis in western Tasmanides: Insights from reactivated basement structures and formation of an ocean-continent transform boundary off western Tasmania. *Gond. Res.,* 19, 608-627.

GIBSON, G.M., TOTTERDELL, J.M., WHITE, L.T., MITCHELL, C.M., STACEY, A.R., MORSE, M.P. & WHITAKER, A. (2013) Pre-existing basement structure and its influence on continental rifting and fracture zone development along Australia's southern rifted margin. *J. Geol. Soc.,* 170, 365–377.

GÖLKE, M. & COBLENTZ, D. (1996) Origins of the European regional stress field. *Tectonophys.,* 266, 11-24.

GÓMEZ, M. & VERGÉS, J. (2005) Quantifying the contribution of tectonics vs. differential compaction in the development of domes along the Mid-Norwegian Atlantic margin. *Basin Research*, 17, 289-310.

GREEN, P.F., CROWHURST, P.V. & DUDDY, I.R. (2004) Integration of AFTA and (U-Th)/He thermochronology to enhance the resolution and precision of thermal history reconstruction in the Anglesea-1 well, Otway Basin, SE Australia. In: *Eastern Australian Basins Symposium II* (Ed. by P.J. Boult, D.R. Johns and S.C. Lang), *Petrol. Explor. Soc. Aust., Spec. Publ.*, 117-131.

HARROWFIELD, M. & KEEP, M. (2005) Tectonic modification of the Australian North-West Shelf: episodic rejuvenation of long-lived basin divisions. *Basin Research*, 17, 225–239.

HILL, K.C. & HALL, R. (2003) Mesozoic-Cenozoic evolution of Australia's New Guinea margin in a West Pacific context. In: *Evolution and Dynamics of the Australian Plate* (Ed. by R.R. Hillis & D. Müller), *Geol. Soc. Aust., Spec. Publ* , 22, 265-290.

HILL, K.C., HILL, K.A., COOPER, G.T., O'SULLIVAN, A.J., O'SULLIVAN, P.B. & RICHARDSON, M.J. (1995) Inversion around the Bass Basin, SE Australia. In: *Basin Inversion* (Ed. by J.G. Buchanan & P.G. Buchanan), *Geol. Soc. Spec. Publ.*, 88, 525-547.

HILLIS, R.R. & REYNOLDS, S.D. (2000) The Australian stress map. *J. Geol. Soc.*, 157, 915–921.

HILLIS, R.R., MONTE, S.A., TAN, C.P. & WILLOUGHBY, D.R. (1995) The contemporary stress field of the Otway Basin, South Australia: Implications for hydrocarbon exploration and production. *APEA J.*, 35, 494-506.

HILLIS, R.R., HOLFORD, S.P., GREEN, P.F., DORÉ, A.G., GATLIFF. R.W., STOKER, M.S., THOMSON, K., TURNER, J.P., UNDERHILL, J.R. & WILLIAMS, G.A. (2008a) Cenozoic exhumation of the southern British Isles. *Geology*, 36, 371-374.

HILLIS, R.R., SANDIFORD, M., REYNOLDS, S.D. & QUIGLEY, M.C. (2008B) Present-day stress, seismicity and Neogene-to-Recent tectonics of Australia's 'passive' margins:

intraplate deformation controlled by plate boundary forces. In: *The Nature and Origin of Compression in Passive Margins* (Ed. by H. Johnson, A.G. Doré, R.W. Gatliff, R. Holdsworth, E.R. Lundin, and J.D. Ritchie), *Geol. Soc. Spec. Publ.*, 306, 71-90.

HOLDGATE, G. R. & GALLAGHER, S. J. (2003) Tertiary, a period of transition to marine basin environments. In: *Geology of Victoria* (Ed. by W.D. Birch), *Geol. Soc. Aust., Spec. Publ.*, 23, 289-335.

HOLDGATE, G.R., SMITH, T.A.G., GALLAGHER, S.J. & WALLACE, M.W. (2001) Geology of coal-bearing Palaeogene sediments, onshore Torquay Basin, Victoria. *Aus. J. Earth Sci.*, 48, 657-679.

HOLFORD, S.P., GREEN, P.F., TURNER, J.P., WILLIAMS, G.A, HILLIS, R.R., TAPPIN, D.R. & DUDDY, I.R. (2008) Evidence for km-scale Neogene exhumation driven by compressional deformation in the Irish Sea basin system. In: *The Nature and Origin of Compression in Passive Margins* (Ed. by H. Johnson, A.G. Doré, R.W. Gatliff, R. Holdsworth, E.R. Lundin, and J.D. Ritchie), *Geol. Soc. Spec. Publ.*, 306, 91-119..

HOLFORD, S.P., GREEN, P.F., DUDDY, I.R., TURNER, J.P., HILLIS, R.R. & STOKER, M.S. (2009a) Regional intraplate exhumation episodes related to plate boundary deformation. *Geol. Soc. Am., Bull.*, 121, 1611-1628.

HOLFORD S.P., TURNER J.P., GREEN P.F. & HILLIS R.R. (2009b) Signature of cryptic sedimentary basin inversion revealed by shale compaction data in the Irish Sea, western British Isles. *Tectonics*, 28, TC4011, doi:10.1029/2008TC002359.

HOLFORD, S.P., GREEN, P.F., HILLIS, R.R., UNDERHILL, J.R., STOKER, M.S. & DUDDY, I.R. (2010a) Multiple post-Caledonian exhumation episodes across northwest Scotland revealed by apatite fission track analysis. *J. Geol. Soc.*, 167, 675–694.

HOLFORD, S.P., HILLIS, R.R., DUDDY, I.R., GREEN, P.F., TUITT, A.K. & STOKER, M.S. (2010b) Impacts of Neogene-Recent compressional deformation and uplift on hydrocarbon prospectivity of the 'passive' southern Australian margin. *APPEA J.*, 50, 267-284.

HOLFORD, S.P., HILLIS, R.R., DUDDY, I.R., GREEN, P.F., STOKER, M.S., TUITT, A.K., BACKE, G., TASSONE, D.R. & MACDONALD, J.D. (2011a) Cenozoic post-breakup compressional deformation and exhumation of the southern Australian margin. *APPEA J.*, 51, 613-638.

HOLFORD, S.P., HILLIS, R.R., DUDDY, I.R., GREEN, P.F., TASSONE, D.R. & STOKER, M.S. (2011b). Palaeothermal and seismic constraints on late Miocene-Pliocene uplift and deformation in the Torquay sub-basin, southern Australian margin. *Aust. J. Earth Sci.*, 58, 543-562.

HOLFORD, S.P., HILLIS, R.R., HAND, M. & SANDIFORD, M. (2011c) Thermal weakening localizes intraplate deformation along the southern Australian continental margin. *Earth Planet. Sci. Lett.*, 305, 217-214

HOLFORD, S.P., SCHOFIELD, N., MACDONALD, J.D., DUDDY, I.R. & GREEN, P.F. (2012) Seismic analysis of igneous systems in sedimentary basins and their impacts on hydrocarbon prospectivity: examples from the southern Australian margin. *APPEA J.*, 52, 229-252.

HUDEC, M.R. & JACKSON, M.P.A. (2002) Structural segmentation, inversion and salt tectonics on a passive margin: Evolution of the Inner Kwanza Basin, Angola. *Geol. Soc. Am., Bull.*, 114, 1222-1244.

JACKSON, C. A.-L. (2012) Seismic reflection imaging and controls on the preservation of ancient sill-fed magmatic vents. *J. Geol. Soc.*, 169, 503–506.

JENSEN-SCHMIDT, B., COCKSHELL, C.D. & BOULT, P.J. (2002) Chapter 5: Structural and tectonic setting. In: *Petroleum Geology of South Australia. Vol. 1: Otway Basin, Second Edition.* (Ed. by P.J. Boult and J.E. Hibburt).

JOHNSON, H., DORÉ, A.G., GATLIFF, R.W., HOLDSWORTH, R., LUNDIN, E.R. & RITCHIE, J.D. (Eds) (2008) *The Nature and Origin of Compression in Passive Margins*, Special Publications, 306, Geological Society, London.

KENNETT, B.L.N., SALMON, M., SAYGIN, E. & AUSMOHO WORKING GROUP (2011) AusMoho: the variation of Moho depth in Australia. *Geophs. J. Int.*, 187, 946-958.

KING, R., HOLFORD, S., HILLIS, R., TUITT, A., SWIERCZEK, E., BACKÉ, G., TASSONE, D. & TINGAY, M (2012) Reassessing the in-situ stress regimes of Australia's petroleum basins. *APPEA J.*, 52, 415-425.

KRASSAY, A.A., CATHRO, D.L. & RYAN, D.J. (2004) A regional tectonostratigraphic framework for the Otway Basin. In: *Eastern Australian Basins Symposium II* (Ed. by P.J. Boult, D.R. Johns and S.C. Lang), *Petrol. Explor. Soc. Aust., Spec. Publ.,* 97-106.

KRISHNA, K.S., BULL, J.M. & SCRUTTON, R.A. (2001) Evidence for multiphase folding of the central Indian Ocean lithosphere. *Geology*, 29, 715-718.

KULPECZ, A.A., MILLER, K.G., BROWNING, J.V., EDWARDS, L.E., POWARS, D.S., MCLAUGHLIN, P.P., JR., HARRIS, A.D. & FEIGENSON, M.D. (2009) Post-impact deposition in the Chesapeake Bay impact structure: Variations in eustasy, compaction, sediment supply, and passive-aggressive tectonism. In: *The ICDP-USGS Deep Drilling Project in the Chesapeake Bay Impact Structure: Results from the Eyreville Core Holes* (Ed. by G.S. Gohn, C. Koeberl, K.G. Miller, and W.U. Reimold), *Geol. Soc. Am., Spec. Paper,* 458, 811-837.

LE BRETON, E., COBBOLD, P.R., DAUTEUIL, O. & LEWIS, G. (2012) Variations in amount and direction of seafloor spreading along the northeast Atlantic Ocean and resulting deformation of the continental margin of northwest Europe. *Tectonics,* 31, TC5006, doi:10.1029/2011TC003087.

LEONARD, M. & CLARK, D. (2011) A record of stable continental region earthquakes from Western Australia spanning the late Pleistocene: Insights for contemporary seismicity. *Earth Planet. Sci. Lett.,* doi:10.1016/j.epsl.2011.06.035.

LI, Q., JAMES, N. P. & MCGOWRAN, B. (2003) Middle and Late Eocene Great Australian Bight lithobiostratigraphy and stepwise evolution of the southern Australian continental margin. *Aust. J. Earth Sci.,* 50, 113-128.

LIU, M., STEIN, S. & WANG, H. (2011) 2000 years of migrating earthquakes in North China: How earthquakes in midcontinents differ from those at plate boundaries. *Lithosphere,* 3, 128-132.

LUNDIN, E.R. & DORÉ, A.G. (2002) Mid-Cenozoic post-breakup deformation in the 'passive' margins bordering the Norwegian-Greenland Sea. *Mar. Petr. Geology,* 19, 79-93.

LUNDIN, E.R. & DORÉ, A.G. (2011) Hyperextension, serpentinization, and weakening: A new paradigm for rifted margin compressional deformation. *Geology,* 39, 347-350.

MACDONALD, J.D., HOLFORD, S.P., GREEN, P.F., DUDDY, I.R., KING, R.C. & BACKÉ, G. (2013) Detrital zircon data reveal the origin of Australia's largest delta system. *J. Geol. Soc.,* 170, 3–6.

MCCLAY, K.R. (1995) The geometries and kinematics of inverted fault systems: A review of analogue model studies. In: *Basin Inversion* (Ed. by J.G. Buchanan & P.G. Buchanan), *Geol. Soc. Spec. Publ.*, 88, 97-118.

MCGOWRAN, B., HOLDGATE, G. R., LI, Q. & GALLAGHER, S. J. (2004) Cenozoic stratigraphic succession in southeastern Australia. *Aust. J. Earth Sci.,* 51, 459-496.

MESSENT, B.E., COLLINS, G.I. & WEST, B.G. (1999) Hydrocarbon prospectivity of the offshore Torquay sub-basin, Victoria: gazettal area V99-1. *Victoria Initiative for Minerals and Petroleum*, 60.

MILLER, J.McL., NORVICK, M.S. & WILSON, C.J.L. (2002). Basement controls on rifting and the associated formation of ocean transform faults–Cretaceous continental extension of the southern margin of Australia. *Tectonophysics,* 359, 131-155.

MITCHUM, R. M., VAIL, P. R. & THOMPSON, S. (1977) Seismic stratigraphy and global changes of sea level, Part 6: stratigraphic interpretation of seismic reflection patterns in depositional sequences. In: *Seismic Stratigraphy - Applications to Hydrocarbon Exploration* (Ed. by C.E. Payton), *AAPG Mem.,* 26, 117-133.

MOSAR, J., LEWIS, G. & TORSVIK, T.H. (2002) North Atlantic sea-floor spreading rates: implications for the Tertiary development of inversion structures of the Norwegian-Greenland Sea. *J. Geol. Soc.,* 159, 503-515.

MÜLLER, R.D., DYKSTERHUIS, S. & REY, P. (2012) Australian paleo-stress fields and tectonic reactivation over the past 100 Ma. *Aust. J. Earth Sci.,* 59, 13-28.

NELSON, E.J. & HILLIS, R.R. (2005) In situ stresses of the West Tuna area, Gippsland Basin. *Austr. J. Earth Sci*., 52, 299-313.

NELSON, E.J., HILLIS, R.R., SANDIFORD, M., REYNOLDS, S.D. & MILDREN, S.D. (2006) Present-day state-of-stress of southeast Australia. *APPEA J.,* 46, 283–305.

NORVICK, M.S. & SMITH, M.A. (2001) Mapping the plate tectonic reconstruction of southern and southeastern Australia and implications for petroleum systems. *APPEA J.*, 41, 15–35.

PASCAL, C., & CLOETINGH, S.A.P.L. (2009) Gravitational potential stresses and stress field of passive continental margins: Insights from the south-Norway shelf. *Earth Planet. Sci. Lett.*, 277, 464-473.

PATRIAT, P. & ACHACHE, J. (1984) India–Eurasia collision chronology has implications for crustal shortening and driving mechanism of plates. *Nature*, 311, 615-621.

PERINCEK, D. & COCKSHELL, C.D. (1995) The Otway Basin: Early Cretaceous rifting to Neogene inversion. *APEA J.*, 35, 451–466.

PERINCEK, D., COCKSHELL, C.D., FINLAYSON, D.M. & HILL, K.A. (1994) The Otway Basin: Early Cretaceous rifting to Miocene strike-slip. In: *NGMA/PESA Otway Basin Symp.*,(Ed. by D.M. Finlayson), AGSO record 1994/14, 27–33.

PEREIRA, R., ALVES, T.M. & CARTWRIGHT, J. (2011) Post-rift compression on the SW Iberian margin (eastern North Atlantic): a case for prolonged inversion in the ocean-continent transition zone. *J. Geol. Soc.*, 168, 1249-1263.

PÉRON-PINVIDIC, G., MANATASCHAL, G., DEAN, S.M. & MINSHULL, T.A. (2008) Compressional structures on the West Iberia rifted margin: controls on their distribution. In: *The Nature and Origin of Compression in Passive Margins* (Ed. by H. Johnson, A.G. Doré, R.W. Gatliff, R. Holdsworth, E.R. Lundin, and J.D. Ritchie), *Geol. Soc. Spec. Publ.*,306, 169-183.

POLLOCK, R.M. (2003) *Sequence stratigraphy of the Paleocene to Miocene Gambier Sub-Basin, Southern Australia.* University of Adelaide, Adelaide, Australia.

PRAEG, D., STOKER, M.S., SHANON, P.M., CERAMICOLA, S., HJELSTUN, B.O. & MATHIESEN, A. (2005) Episodic Cenozoic tectonism and the development of the NW European 'passive' continental margin. *Mar. Pet. Geology*, 22, 1007–1030.

QUIGLEY, M., CUPPER, M. & SANDIFORD, M. (2006) Quaternary faults of southern Australia: palaeoseismicity, slip rates and origin. *Aust. J. Earth Sci.*, 53, 285-301

REYNOLDS, S.D., COBLENTZ, D.D. & HILLIS, R.R. (2002) Tectonic forces controlling the regional intraplate stress field in continental Australia: results from new finite-element modelling. *J. Geophys. Res.*, 107, B7, doi:10.1029/2001JB000408.

REYNOLDS, S.D., COBLENTZ, D.D. & HILLIS, R.R. (2003) Influences of plate-boundary forces on the regional intraplate stress field of continental Australia. In: *Evolution and Dynamics of the Australian Plate* (Ed. by R.R. Hillis & D. Müller), *Geol. Soc. Aust., Spec. Publ* , 22, 59-70.

REYNOLDS, S.D., MILDREN, S.D., HILLIS, R.R. AND MEYER, J.J. (2006) Constraining stress magnitudes using petroleum exploration data in the Cooper-Eromanga Basins, Australia. *Tectonophys.*, 415, 123-140.

RITCHIE, J.D., JOHNSON, H., QUINN, M.F. & GATLIFF, R.W. (2008) The effects of Cenozoic compression within the Faroe-Shetland Basin and adjacent areas. In: *The Nature and Origin of Compression in Passive Margins* (Ed. by H. Johnson, A.G. Doré, R.W. Gatliff, R. Holdsworth, E.R. Lundin, and J.D. Ritchie), *Geol. Soc. Spec. Publ.*,306, 121-136.

SANDIFORD, M. (2003a) Geomorphic constraints on the late Neogene tectonics of the Otway Range. *Aust. J. Earth Sci.*, 50, 69–80.

SANDIFORD, M. (2003b) Neotectonics of south-eastern Australia: linking the Quaternary faulting record with seismicity and in situ stress. In: *Evolution and Dynamics of the Australian Plate* (Ed. by R.R. Hillis & D. Müller), *Geol. Soc. Aust., Spec. Publ* , 22, 107–120.

SANDIFORD, M. & EGHOLM, D.L. (2008) Enhanced intraplate seismicity across continental margins: some causes and consequences. *Tectonophysics*, 457, 197-208.

SANDIFORD, M. & QUIGLEY, M. (2009) TOPO-OZ: Insights into the various modes of intraplate deformation in the Australian continent. *Tectonophys.*, 474, 405-416.

SANDIFORD, M., WALLACE, M. & COBLENTZ, D. (2004) Origin of the in situ stress field in southeastern Australia. *Basin Research*, 16, 325–338.

SCHNEIDER, C.L., HILL, K.C. & HOFFMAN, N. (2004) Compressional growth of the Minerva Anticline, Otway Basin, Southeast Australia-evidence of oblique rifting. *APPEA J.*, 44, 463–80.

STOKER, M.S., HOULT, R.J., NIELSEN, T., HJELSTUN, B.O., LABERG, J.S., SHANNON, P.M., PRAEG, D., MATHIESEN, A., VAN WEERING, T.C.E. & McDONNELL, A. (2005) Sedimentary and oceanographic responses to early Neogene compression on the NW European margin. *Mar. Petr. Geology*, 22, 1031-1044.

STOKER, M.S., HOLFORD, S.P., HILLIS, R.R., GREEN, P.F. & DUDDY, I.R. (2010) Cenozoic post-rift sedimentation off NW Britain: recording the detritus of episodic uplift on a passive continental margin. *Geology*, 39, 595-598.

TASSONE, D.R., HOLFORD, S.P., TINGAY, M.R.P., TUITT, A.K., STOKER, M.S. & HILLIS, R.R. (2011) Overpressures in the central Otway Basin: the result of rapid Pliocene-Recent sedimentation? *APPEA J.*, 51, 439-458.

TASSONE, D.R., HOLFORD, S.P., HILLIS, R.R. & TUITT, A.K. (2012) Quantifying Neogene plate-boundary controlled uplift and deformation of the southern Australian margin. In: *Faulting, Fracturing and Igneous Intrusion in the Earth's Crust* (Ed. by D. Healy, R.W.H. Butler, Z.K. Shipton, and R.H. Sibson), *Geol. Soc. Spec. Publ.*, 367, 91-110.

TASSONE, D.R., HOLFORD, S.P., DUDDY, I.R., GREEN, P.F. & HILLIS, R.R. (2013) Quantifying Cretaceous-Cenozoic exhumation in the Otway Basin using sonic velocity data: implications for conventional and unconventional hydrocarbon prospectivity. *AAPG Bull.*, doi:10.1306/04011312111

TICKELL S. J., EDWARDS J. & ABELE C. (1992) Port Campbell Embayment 1:100 000 map geological report (and map). *Geological Survey of Victoria Report* 95.

TOTTERDELL, J. (2012) New exploration opportunities along Australia’s southern margin. *APPEA J.*, 51, 439-458.

TOWNEND, J. & ZOBACK, M.D. (2000) How faulting keeps the crust strong. *Geology*, 28, 399-402.

TRUPP, M.A., SPENCE, K.W. & GIDDING, M.J. (1994) Hydrocarbon prospectivity of the Torquay Sub-Basin, Offshore Victoria. *APEA J.*, 34, 479–494.

TUITT, A., UNDERHILL, J.R., RITCHIE, J.D., JOHNSON, H. & HITCHEN, K. (2010) Timing, controls and consequences of compression in the Rockall-Faroe area of the NE Atlantic Margin. In: *Proceedings of the 7th Petroleum Geology Conference: From mature basins to new frontiers* (Ed. by B.A. Vining & S. Pickering), pp. 963-977. The Geological Society, London.

TURNER, J.P. & WILLIAMS, G.D. (2004) Sedimentary basin inversion and intra-plate shortening. *Earth Sci. Reviews* 65, 277-304.

TURNER, J.P., GREEN, P.F., HOLFORD, S.P. & LAWRENCE, S.R. (2008) Thermal history of the Rio Muni (West Africa)-NE Brazil margins during continental breakup. *Earth Planet. Sci. Lett.*, 270, 354-367.

VÅGNES, E., GABRIELSEN, R.H. & HAREMO, P. (1998) Late Cretaceous-Cenozoic intraplate contractional deformation at the Norwegian continental shelf: timing, magnitude and regional implications. *Tectonophys.* 300, 29–46.

VÁZQUEZ, J.T., MEDIALDEA, T., ERCILLA, G., SOMOZA, L., ESTRADA, F., FERNÁNDEZ PUGA, M.C., GALLART, J., GRÀCIA, E., MAESTRO, A. & SAYAGO, M. (2008) Cenozoic deformational structures on the Galicia Bank Region (NW Iberian continental margin). *Mar. Geology*, 249, 128–149.

VEEVERS, J.J. (2000) Change of tectono-stratigraphic regime in the Australian plate during the 99 Ma (mid-Cretaceous) and 43 Ma (mid-Eocene) swerves of the Pacific. *Geology,* 28, 47-50.

WESSEL, P. & KROENKE, L.W. (2000) Ontong Java Plateau and late Neogene changes in Pacific plate motion. *J. Geophys. Res.*, 105, 28255–28277.

WILLIAMS, G.D., POWELL, C.M. & COOPER, M.A. (1989) Geometry and kinematics of inversion tectonics. In: *Inversion Tectonics* (Ed. by M.A. Cooper & G.D. Williams), *Geol. Soc. Spec. Publ.,* 44, 3-15.

WILLIAMS, G.A., TURNER, J.P. & HOLFORD, S.P. (2005) Inversion and exhumation of the St George’s Channel Basin, offshore Wales, UK. *J. Geol. Soc.*, 162, 97-110.

WILLIAMS, S.E., WHITTAKER, J.M. & MÜLLER, R.D. (2011) Full-fit, palinspastic reconstruction of the conjugate Australian-Antarctic margins. *Tectonics*, 30, TC6012, doi:10.1029/2011TC002912.

ZIEGLER, P.A., CLOETINGH, S. & VAN WEES, J.D. (1995) Dynamics of intra-plate compressional deformation: the Alpine foreland and other examples. *Tectonophys.*, 252, 7-59.

ZOBACK, M.L. & ZOBACK, M. (2007) Lithosphere Stress and Deformation. In: *Treatise on Geophysics* (Ed. by G. Schubert), 6, 253-273.

FIGURE CAPTIONS

Figure 1. Indo-Australian plate boundaries and associated forces, and state of stress within the Australian continent (modified after Reynolds *et al.*, 2002). Large black arrows indicate mid-ocean ridge driving forces, small black arrows indicate resisting forces associated with continental collisions and subduction zones, white arrows indicate buoyancy forces resulting from lithospheric density variations, grey arrows indicate mean maximum horizontal stress orientations within the Australian continent (length of bars is proportional to data quality). H, Himalaya; S, Sumatra Trench; J, Java Trench; B, Banda Arc; NG, New Guinea; SM, Solomon Trench; NH, New Hebrides; LHR, Lord Howe Rise; TK, Tonga-Kermadec Trench; NZ, New Zealand; SNZ, south of New Zealand; MOR, mid-ocean ridge; cb, collisional boundary; sz, subduction zone; is, island arc.

Figure 2. A. Otway Basin location map. Inset map shows 2D seismic lines used to map Cenozoic unconformities and compressional structures. Northern limit of Cenozoic marine deposits, Lower Cretaceous inliers and major extensional structures modified from Krassay *et al.* (2004). Location/structural element acronyms: COB, continent-ocean boundary; MH, Merino High; MFZ, Mussel Fault Zone; MSB, Morum sub-basin; NSB, Nelson sub-basin; OR, Otway Ranges; PO, Portland Trough; PT, Penola Trough; SF, Sorell Fault; ST, Shipwreck Trough; TFZ, Tartwaup Fault Zone; TSB, Torquay sub-basin. Well acronyms:

A1, Argonaut-1; AS1, Anglesea-1; BB1, Bridgewater Bay-1; BR1, Breaksea Reef-1; CH1, Chama-1A; CM1, Champion-1; CO1, Copa-1; CR1, Crayfish-A1; D1, Digby-1; DB1, Discovery Bay-1; ER1, Eric the Red-1; HC1, Hindhaugh Creek-1; LA1, Loch Ard-1; LB1, La Bella-1; MI1, Minerva-1; MO1, Morum-1; MU1, Mussel-1; NA1, Nautilus-1; NE1, Neptune-1; NO1, Normanby-1; NP1, Neptune-1; PE1, Pecten-1; TI1, Triton-1; TO1, Troas-1; TR1, Trumpet-1; V1, Voluta-1. **B.** Western Otway Basin location map, showing positions of seismic lines displayed in Figures 6-9, and hydrocarbon fields in the western Otway Basin. Cenozoic structural elements based on this study and Perineck & Cockshell (1995). A, B and C refer to structural components of the Copa Anticline. **C.** Eastern Otway Basin location map, showing positions of seismic lines and cross sections displayed in Figures 10-14, and onshore and offshore hydrocarbon fields in the Port Campbell Embayment and Shipwreck Trough. Cenozoic structural elements based on this study, Geary & Reid (1998) and Clark *et al.* (2011). Cenozoic structural element acronyms: BF, Bambra Fault; CCF, Castle Cove Fault; CM, Curdie Monocline; DEM, Devil's Elbow Monocline; Ferguson Hill Anticline, FHA; Johanna Fault. Acronyms of wells not given in Figure 2A: BC1, Barton Corner-1; FH1, Fergusons Hill-1; NP3, North Paaratte-3; PC4, Port Campbell-4; R1, Rowans-1; S1, Seaview-1. **D.** Tectonic elements map for southeastern Australia, showing major basement terranes and structures and location of the continent-ocean boundary (modified after Gibson *et al.*, 2011) and the depth to the seismological Moho (modified after Kennett *et al.*, 2011).

Figure 3. Stratigraphic chart for the Otway Basin (modified after Krassay *et al.*, 2004). Acronyms of regional unconformities used in seismic mapping: MPU, late Miocene-Pliocene unconformity; IOU, intra-Oligocene unconformity; ILU, intra-Lutetian unconformity; IMU, intra-Maastrichtian unconformity; MCU, mid-Cretaceous unconformity.

Figure 4. Lithostratigraphic well correlation from Morum-1 in the northwest part of the Otway Basin to Eric the Red-1 in the southeast, demonstrating variations in thickness of the unconformity-bound Cenozoic Wangerrip, Nirranda and Heytesbury Groups (colour schemes and acronyms are the same as those used in Figure 3). The thickness of the Cenozoic succession is generally less than 1 km, with pronounced thinning across the Morum-Copa and Loch Ard structural highs. Thick localised post-MPU successions in the basin centre represent Pliocene-Recent marine clastic sediments deposited in submarine canyons (Tassone *et al.*, 2011). Distance between wells is not to scale. Formation tops picked using well completion reports and PGS' SAMDA (Southern Australian Margin Digital Atlas) database.

Figure 5. Time-structure maps (in seconds (s) two-way travel-time) based on regional seismic mapping of the **A.** intra-Oligocene, **B.** intra-Lutetian, and **C.** intra-Maastrichtian unconformities.

Figure 6. A. Uninterpreted and **B.** interpreted seismic profile 85-02, perpendicular to the fold axis of the Morum Anticline. Folding of the IMU and ILU is most likely related to reverse-slip along underlying reactivated Cretaceous normal faults, though these faults still exhibit net-normal displacements. Thermal history data from the Morum-1, which indicate substantial mid-Eocene erosion at the ILU are presented in Figure 17A. Seismic profile provided by PGS.

Figure 7. A. Uninterpreted and **B.** interpreted seismic profile 85-13, parallel to the fold axis of the Morum Anticline. Interpretation modified after Duddy *et al.* (2003). Seismic profile provided by PGS.

Figure 8. A. Uninterpreted and **B.** interpreted seismic profile CO88-11, perpendicular to the fold axis of the Copa A anticline. Seismic profile provided by PGS.

Figure 8. A. Uninterpreted and **B.** interpreted seismic profile CO88-11, perpendicular to the fold axis of the Copa A anticline. Folding of the IMU and ILU is most likely related to reverse-slip along underlying reactivated Cretaceous normal faults, though these faults still exhibit net-normal displacements. Seismic profile provided by PGS.

Figure 9. A. Uninterpreted and **B.** interpreted seismic profile CO88-22, perpendicular to the fold axis of the Copa B anticline. Seismic profile provided by PGS.

Figure 10. A. Uninterpreted and **B.** interpreted seismic profile OH91-113, perpendicular to the fold axis of the low-amplitude Pecten Anticline, which correlates with the onshore Curdie Monocline. Black arrows indicate erosional truncation of strata beneath the MPU. Further low-amplitude folding of Cenozoic sediments towards the SSE section of the profile is associated with the Minerva, Point Ronald and Crowes anticlines. Seismic profile provided by DPI Victoria.

Figure 11. Onshore NW-SE cross-section from the Port Campbell Embayment to Otway Ranges, modified after Edwards *et al.* (1996). It is suggested here that the Curdie Monocline, Ferguson Hill Anticline and Crowes Anticline correlate with offshore Cenozoic folds of similar trends (the Pecten, Minerva and Crowes Anticlines, respectively). Note that this interpretation shows (often substantial) net-reverse slip along the shallow segments of many of the normal faults that controlled deposition of the Cretaceous section. Similar evidence for net-reverse displacement on faults imaged on offshore seismic sections is lacking, but this

may reflect increasing intensity of deformation from the SW to NE in the eastern Otway Basin, consistent with the marked increase in onshore topography towards the Otway Ranges.

Figure 12. A. Uninterpreted and **B.** interpreted seismic profile OE80A-1056, parallel to the fold axis of the Minerva Anticline. Thermal history data from the Minerva-1 well are presented in Figure 17C. Small black arrows indicate onlap of late Oligocene-early Miocene sediments onto the crest of the Minerva Anticline. Seismic profile provided by PGS.

Figure 13. A. Uninterpreted and **B.** interpreted seismic profile OE81-2011, perpendicular to the fold axis of the Point Ronald Anticline. Whilst it is possible to interpret key unconformities, poor seismic quality within the Cretaceous section makes identification of faults difficult. This fold is tentatively interpreted as being caused by reverse-reactivation of an E-dipping Cretaceous normal fault. Seismic profile provided by PGS.

Figure 14. A. Uninterpreted and **B.** interpreted seismic profile OH91-210, perpendicular to the fold axes of the Crowes (left) and Loch Ard (right) anticlines. Seismic profile provided by DPI Victoria.

Figure 15. Compilation map showing Cenozoic structures in the Otway Basin (based on this study, Perineck & Cockshell (1995) and Clark *et al.,* (2011)) and present-day maximum horizontal stress orientations determined for a number of onshore and offshore hydrocarbon exploration wells and boreholes (modified after Nelson *et al.,* 2006). Wherever possible, structures have been coded to denote the broad period of time during which their post-breakup activity initiated. Note that it is difficult to demonstrate the pre-late Miocene deformation histories of many of the onshore structures shown on this map, and it is possible

that some of these structures also record earlier deformation. Structures broadly trend ~NE-SW irrespective of the timing of their initiation. These orientations are approximately orthogonal to the ~NW-SE present-day maximum horizontal stress orientation. The focal mechanism for the December 2nd 1977 M_L = 4.4 Balliang earthquake is also shown (after Denham *et al.*, 1981). The focal mechanism is considered to be poorly constrained but nonetheless indicates thrust faulting due to ~NW-SE-directed compression, consistent with present-day stress measurements and geological evidence from this area.

Figure 16. Chronology of post-breakup deformation in the Otway Basin, in relation to regional stratigraphy (after Holdgate & Gallagher, 2003) and plate boundary events that likely influenced the local stress field. Vertical black bars indicate approximate extent of activity of individual structures, based on seismic mapping, stratigraphic evidence and previous studies as described in the text. Vertical white boxes represent estimates of the time intervals during which Cenozoic and/or Cretaceous sedimentary sections began to cool from palaeotemperature peaks based on AFTA data from exploration wells. In most cases the cooling can be attributed to exhumation which accompanied compressional growth of the structures into which wells have been drilled. Post-late Miocene approximate slip rates are indicated for selected onshore faults, after Clark *et al.* (2011). Horizontal grey bars represent time ranges of the unconformities used in seismic mapping. Sources of information used to compile the plate boundary events are as follows: Southern Ocean half-spreading rates (Li *et al.*, 2004); timing of breakup off the Otway Basin and onset of fast-spreading (Norvick & Smith, 2001); Wharton Basin Ridge spreading cessation (Dyksterhuis & Müller, 2008); Tasman Sea spreading cessation (Gaina *et al.*, 1998); orogenic events (Dickinson *et al.*, 2002). Stratigraphic acronyms: BG, Brighton Group; DBG, Demons Bluff Group; EVG,

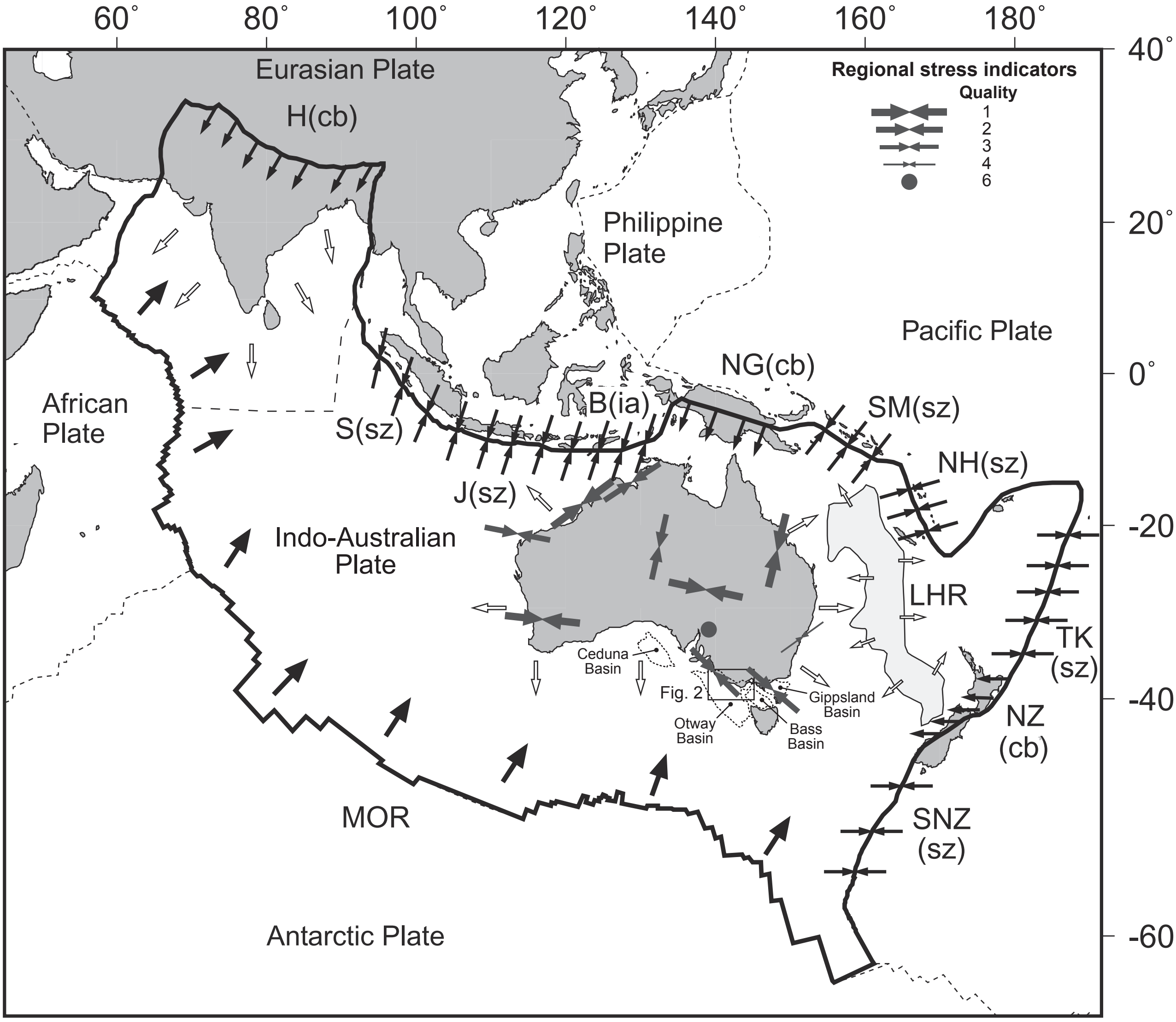
Eastern View Group; HG, Heytesbury Group; NG, Nirranda Group; TG, Torquay Group; WG, Wangerrip Group.

Figure 17. A. (Left) Palaeotemperatures derived from AFTA, VR and apatite (U-Th)/He data from the Morum-1 well plotted against depth and the estimated present-day geothermal gradient. Integration of these data suggests that Cretaceous section began to cool from palaeotemperatures that were significantly higher than present-day temperatures between 57 and 40 Ma. Similarity between the gradients of the present-day and palaeotemperature profiles indicates that cooling was caused by exhumation. Full details of thermal history data and their interpretation are provided by Duddy *et al.* (2003). (Right) Ranges of amounts of section removed from the top-Sherbrook Group unconformity (ILU) and palaeogeothermal gradients (hyperbolic ellipsoid) required to explain the palaeotemperatures estimated from AFTA, apatite (U-Th)/He and VR data. The hyperbolic ellipsoid defines the parameter ranges that are consistent with the palaeotemperature constraints within 95% confidence limits. The black dot indicates maximum likelihood solution values of palaeogeothermal gradient and removed section. For a mid-Eocene palaeogeothermal gradient equal to the present-day geothermal gradient of 29.2°C km^{-1}, 1500 m of section removed at the top Sherbrook Group unconformity (ILU) is allowed by the data. **B.** (Left) Palaeotemperatures derived from AFTA, VR and apatite (U-Th)/He data from the Nerita-1 well plotted against depth and the estimated present-day geothermal gradient. These data indicate that the drilled section has been considerably hotter in the past, and coupled modelling of AFTA and apatite (U-Th)/He data suggest that the Cretaceous-Cenozoic section began to cool from these elevated palaeotemperatures between 10 and 3 Ma. Full details of thermal history data and their interpretation are provided by Holford *et al.* (2011*b*). (Right) Ranges of amounts of section removed from the top-Torquay Group unconformity (MPU) and palaeogeothermal gradients

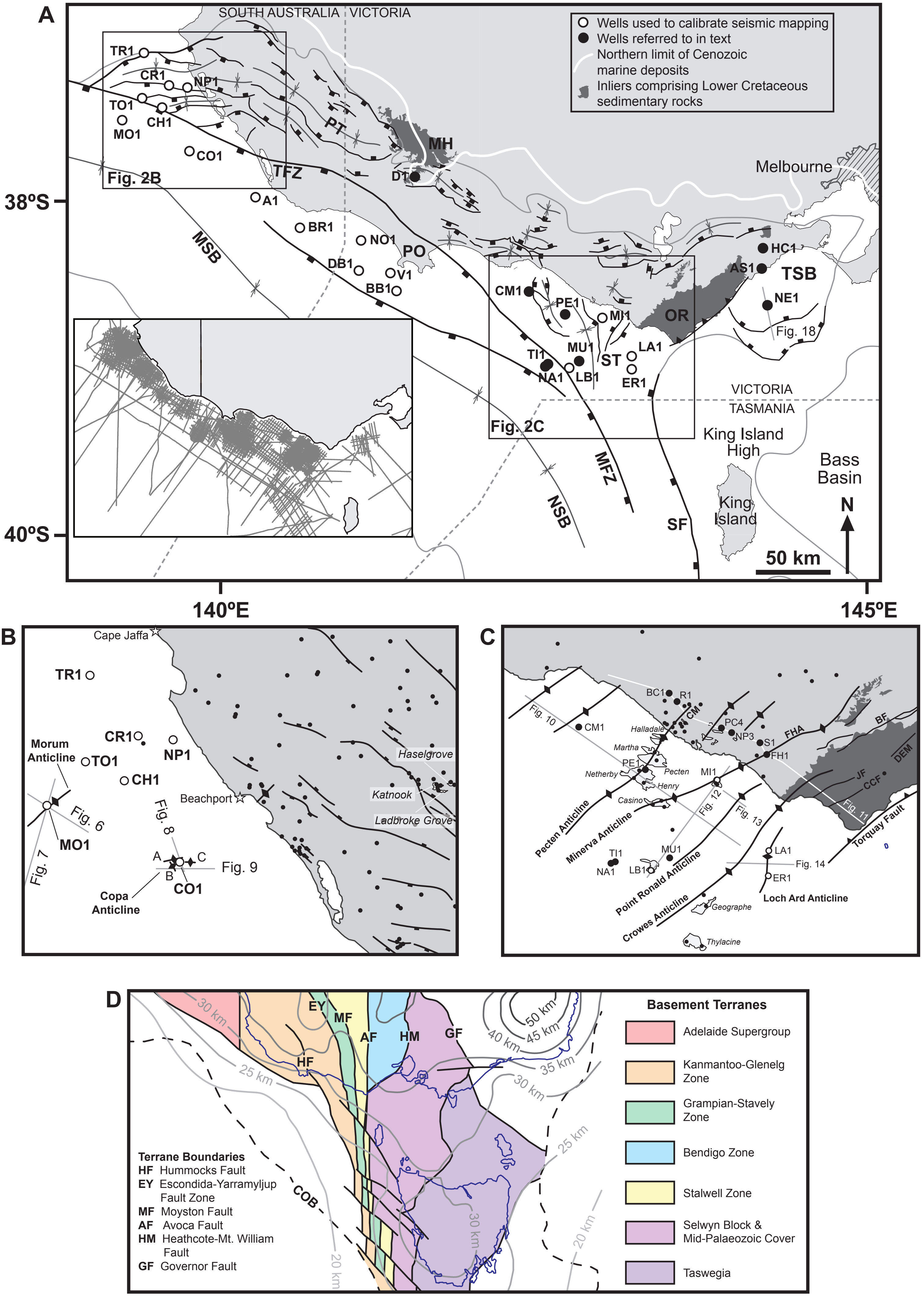
(hyperbolic ellipsoid) required to explain the palaeotemperatures estimated from AFTA, apatite (U-Th)/He and VR data. For a palaeogeothermal gradient of 31.5°C km^{-1}, equal to the present-day gradient, approximately 1100 m of removed section is required on the top-Torquay Group unconformity in order to honour the palaeotemperature constraints. **C.** (Left) Palaeotemperatures derived from AFTA and VR data from the Minerva-1 well plotted against depth and the estimated present-day geothermal gradient. Similarity between the present-day and palaeotemperature estimates suggests that the Cenozoic-Cretaceous section drilled by this well is currently at maximum post-depositional palaeotemperatures. Full details of thermal history data and their interpretation are provided by Holford *et al.* (2010). (Right) Ranges of amounts of allowed section removed from the top-Heytesbury Group unconformity (MPU) and palaeogeothermal gradients (hyperbolic ellipsoid) required to explain the palaeotemperatures estimated from AFTA and VR data. For a palaeogeothermal gradient of 42.2°C km^{-1}, equal to the value of the present-day gradient, no section is required to be removed from the top-Heytesbury Group unconformity to honour constraints from AFTA and VR. About 350 m of erosion is allowed by the data, assuming that the geothermal gradient has increased over time from 35°C km^{-1} at the time of erosion to the present-day value.

Figure 18. A. Uninterpreted and **B.** interpreted seismic profile O40-21, perpendicular to the fold axes of the Nerita Anticline in the Torquay sub-basin. Thermal history data from the Nerita-1 well are presented in Figure 17B. Seismic profile provided by PGS.

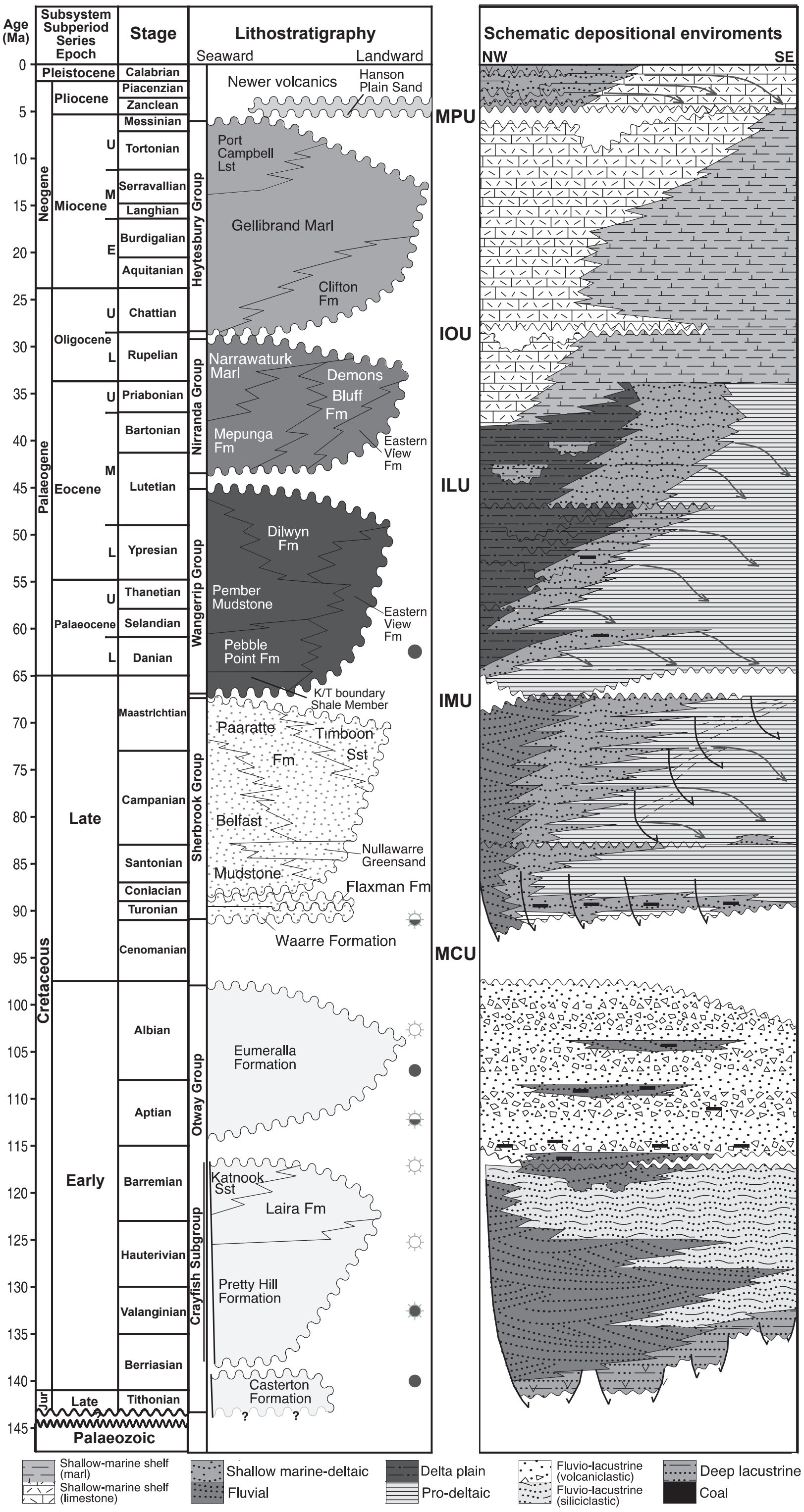
Holford et al Figure 1



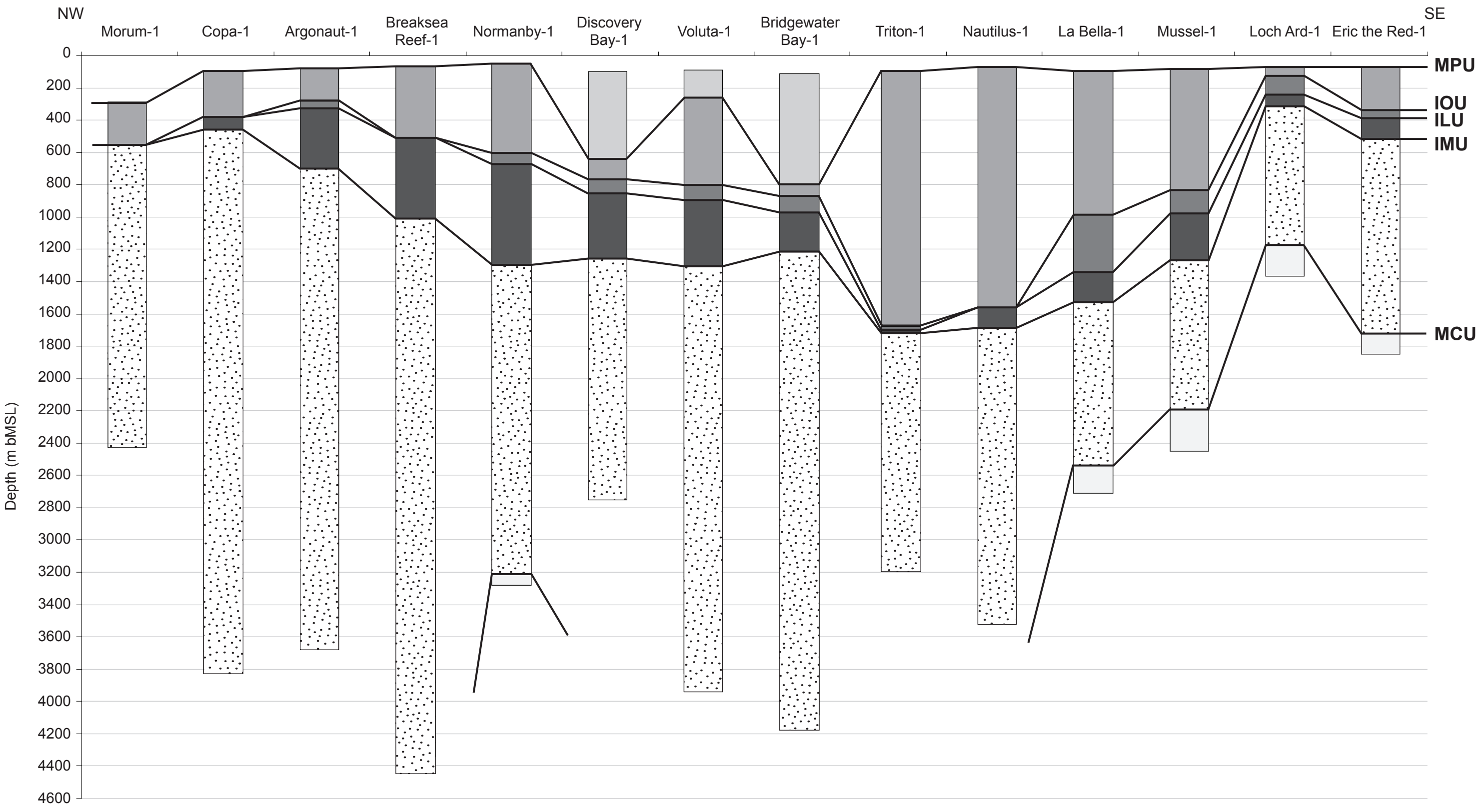
Holford et al Figure 2



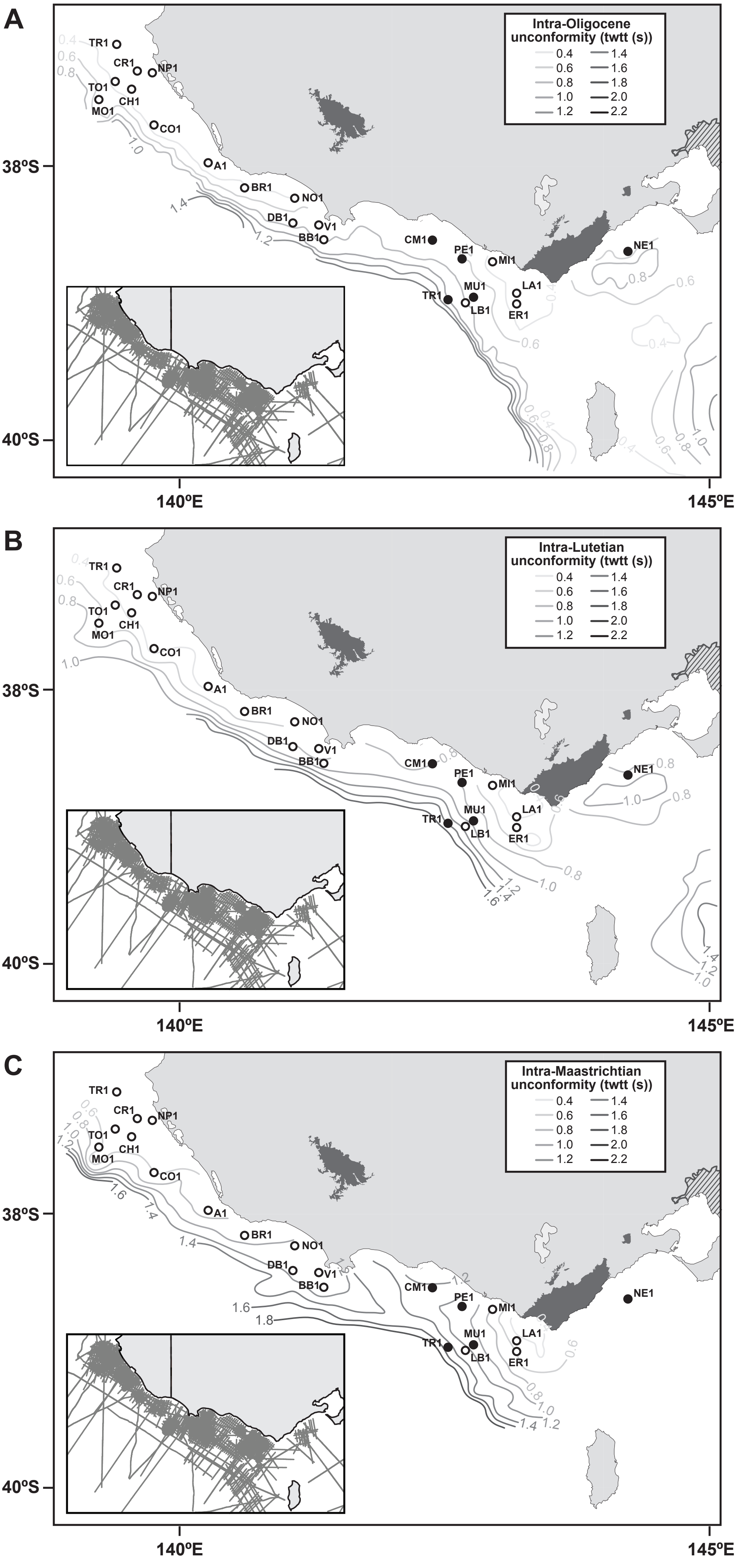
Holford et al Figure 3



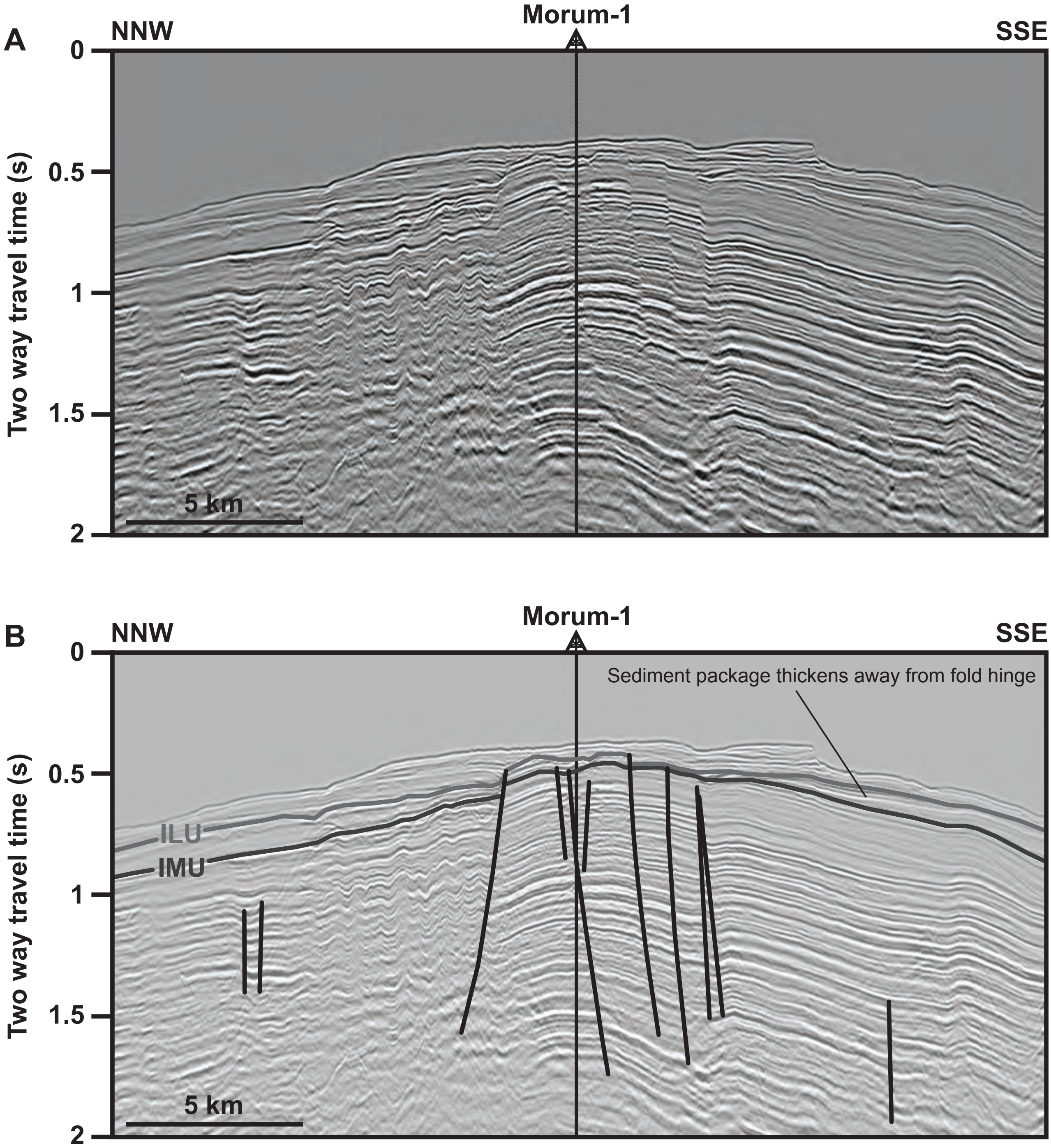
Holford et al Figure 4



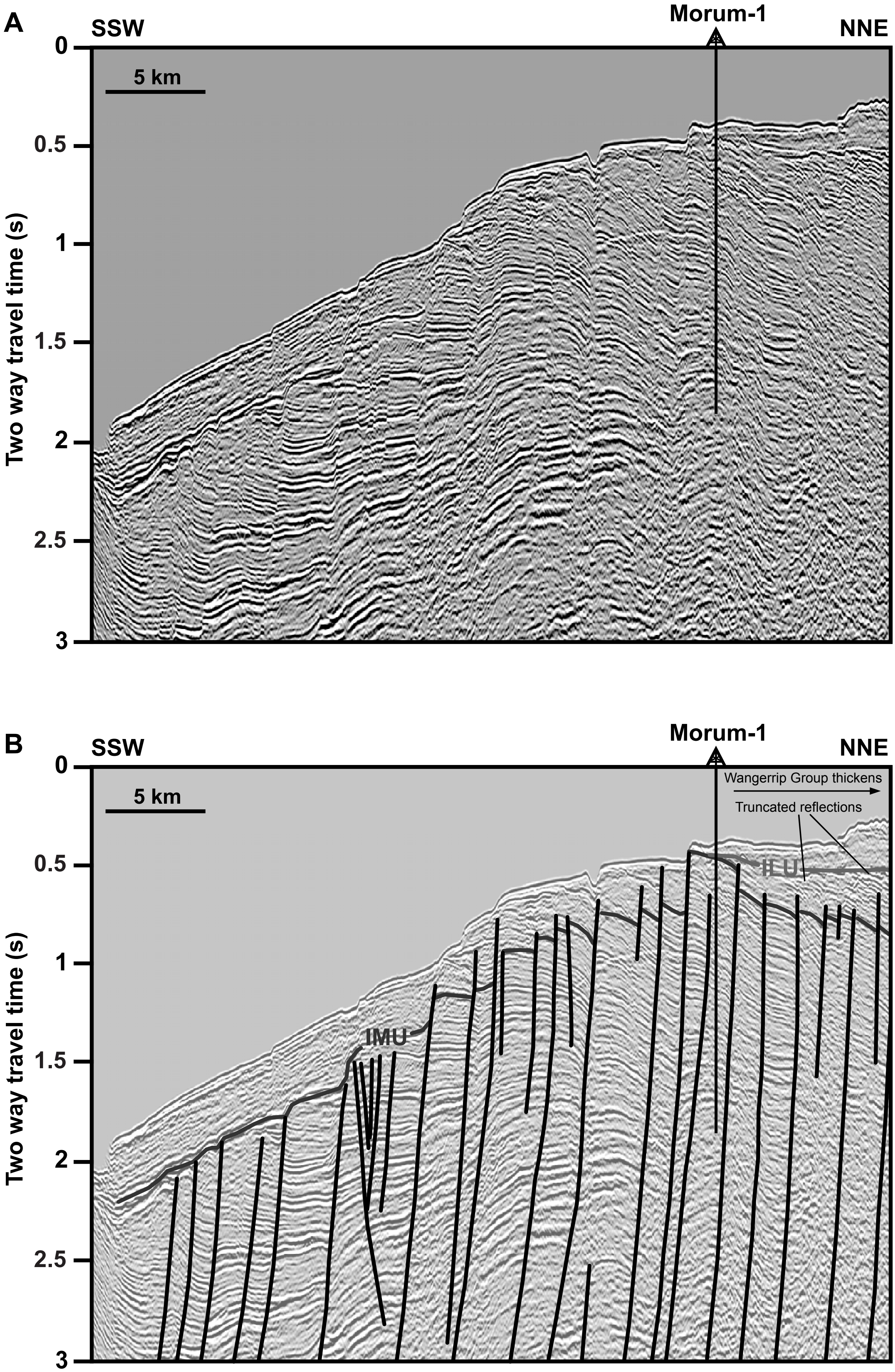
Holford et al Figure 5



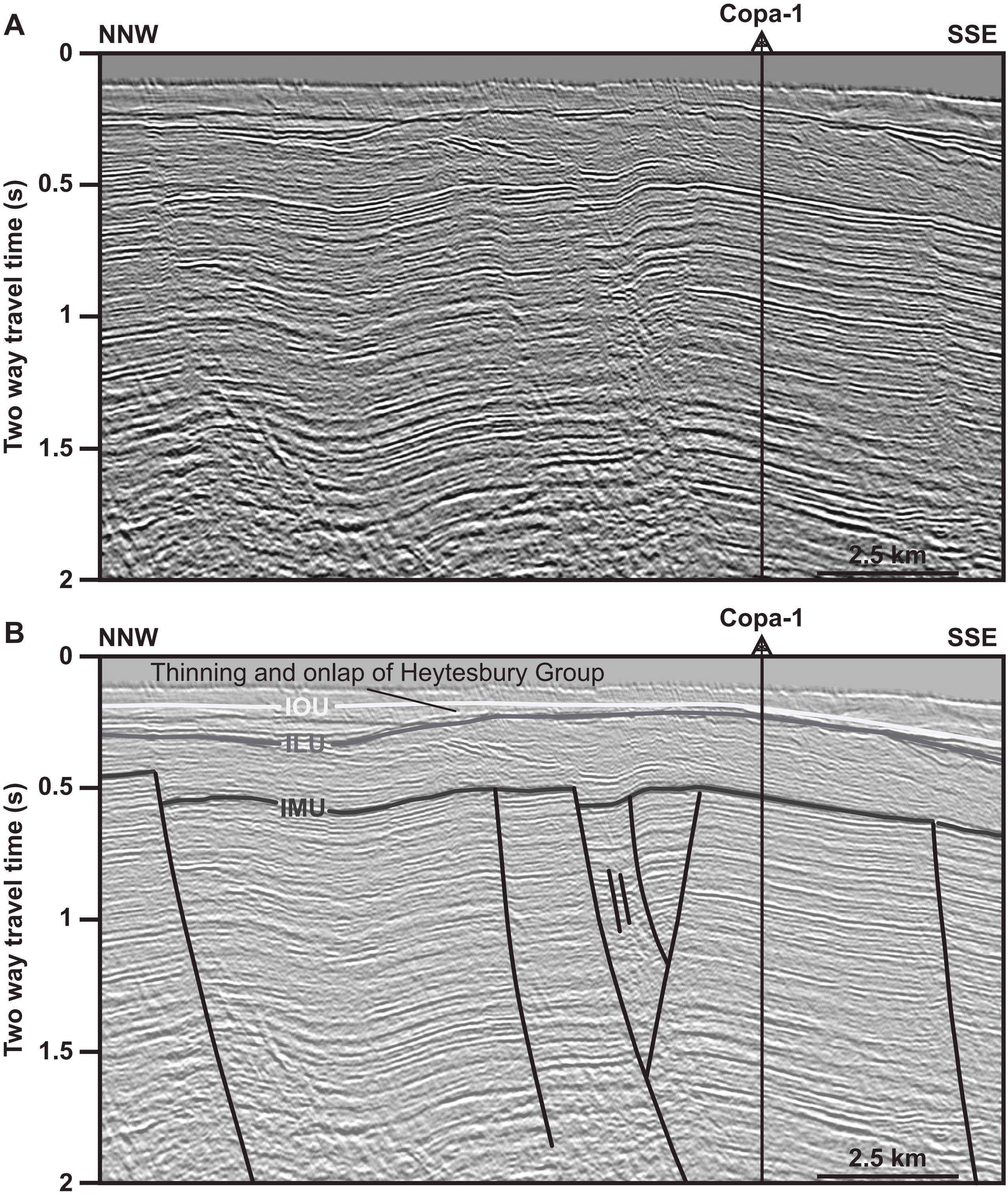
Holford et al Figure 6



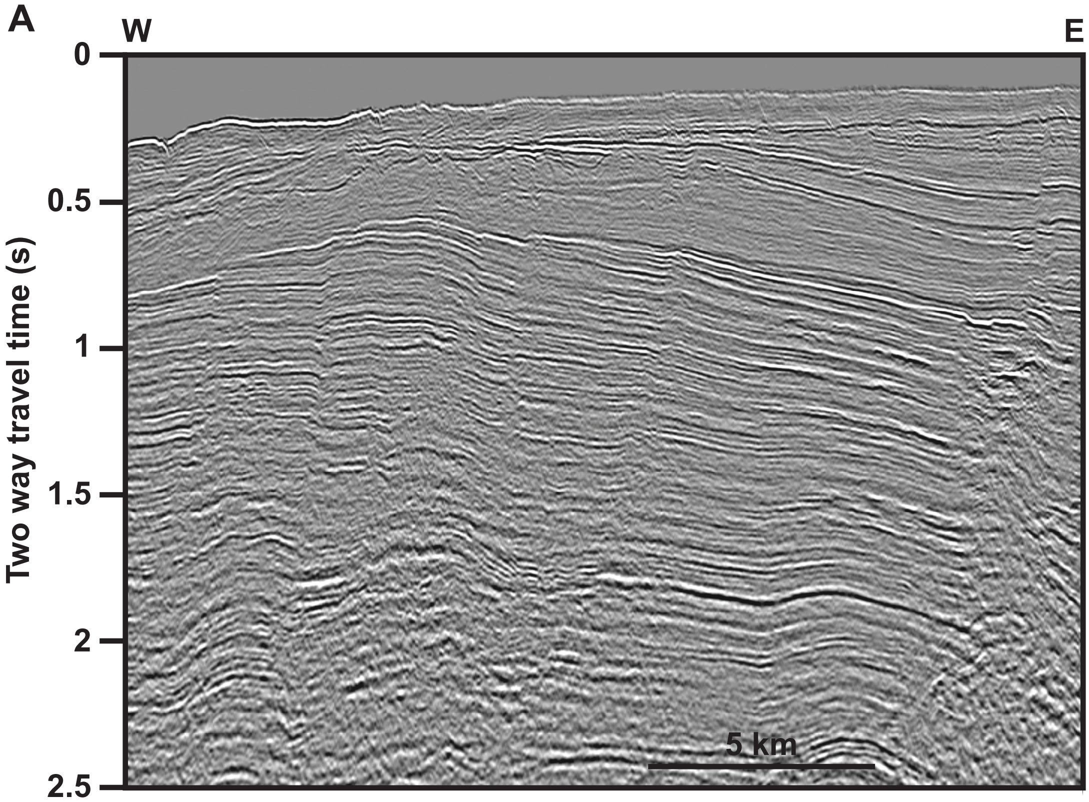
Holford et al Figure 7

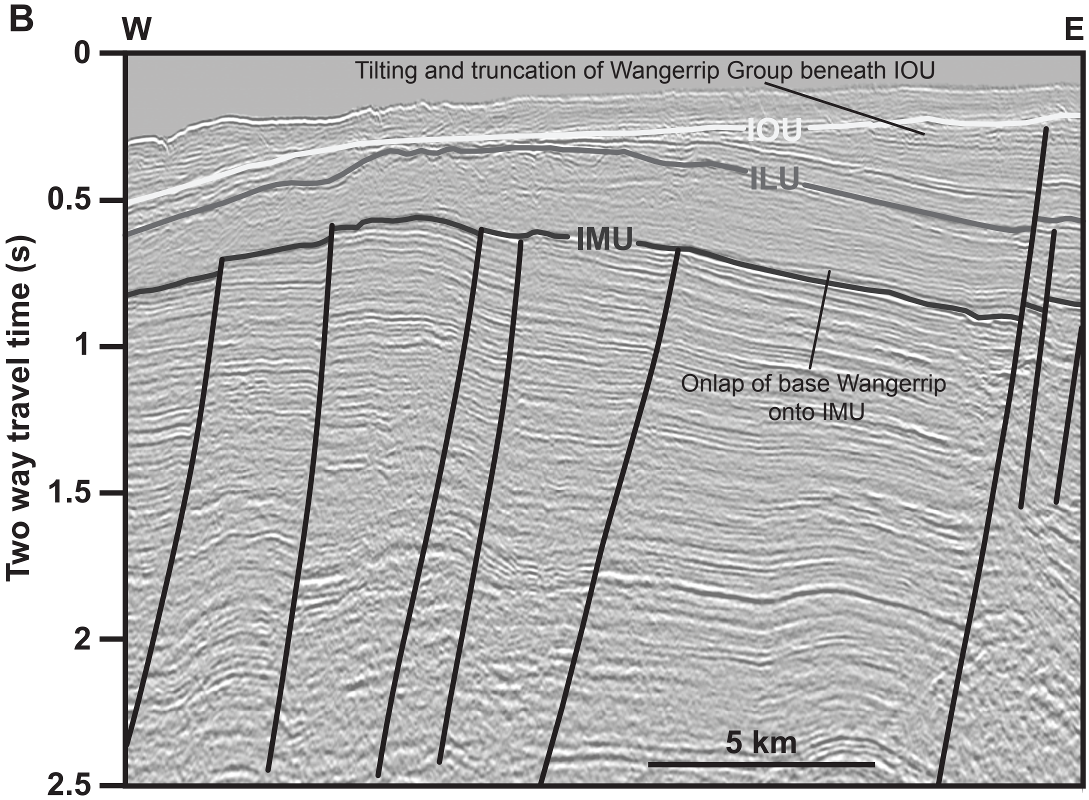


Holford et al Figure 8

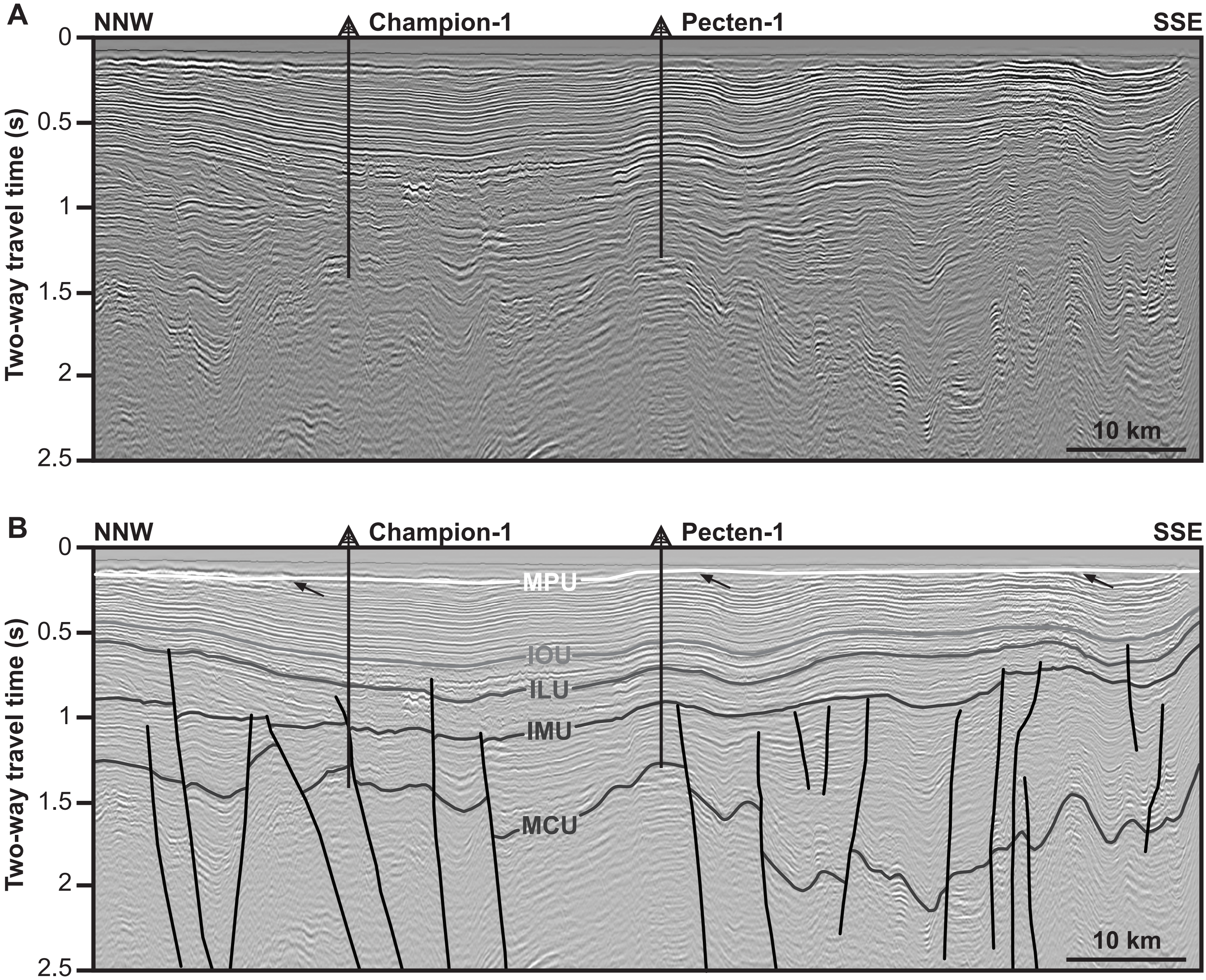


Holford et al Figure 9

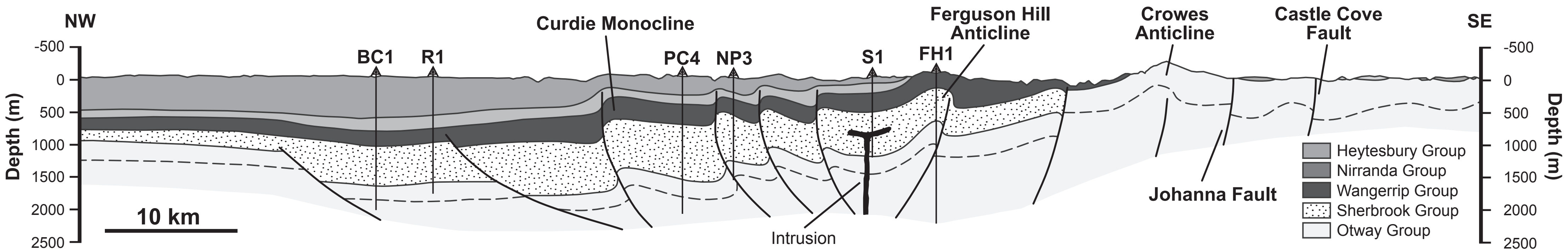




Holford et al Figure 10

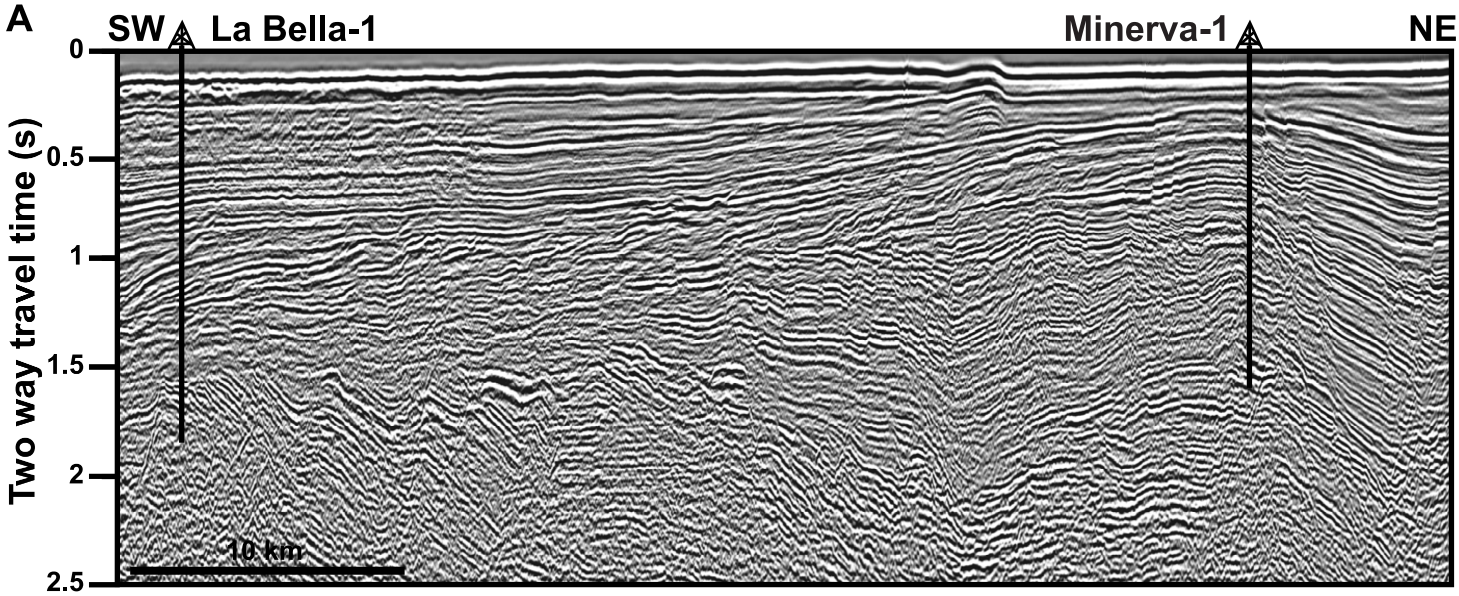


Holford et al Figure 11

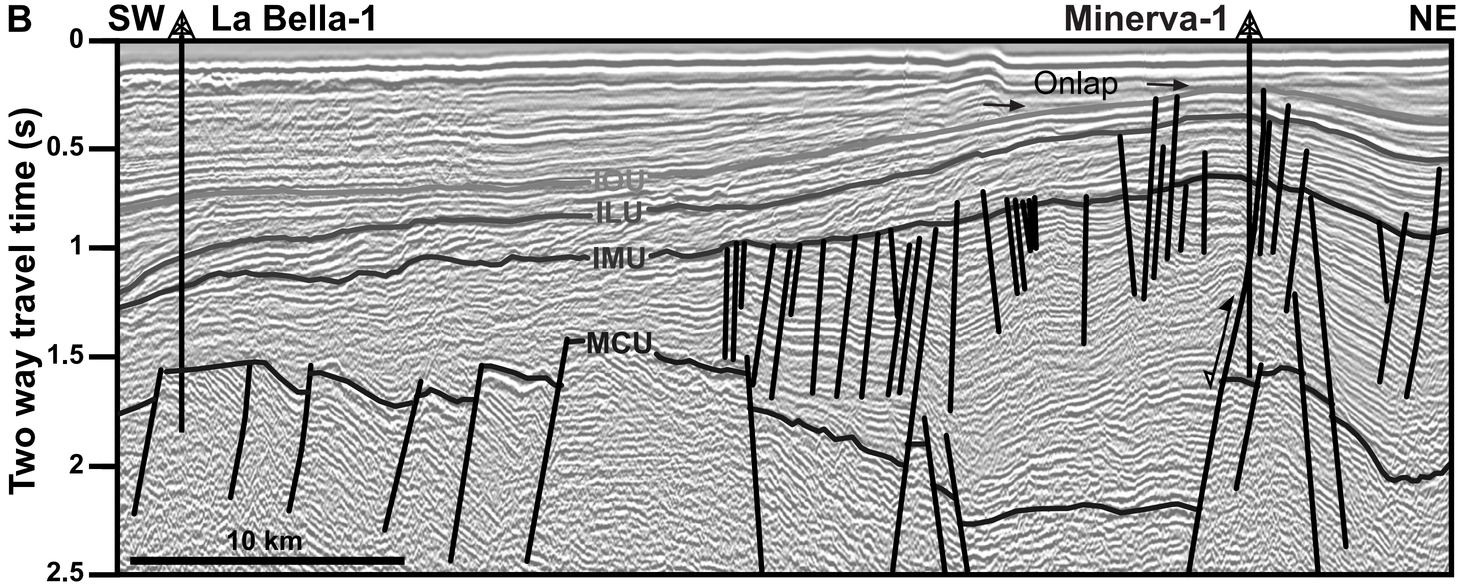


Holford et al Figure 12

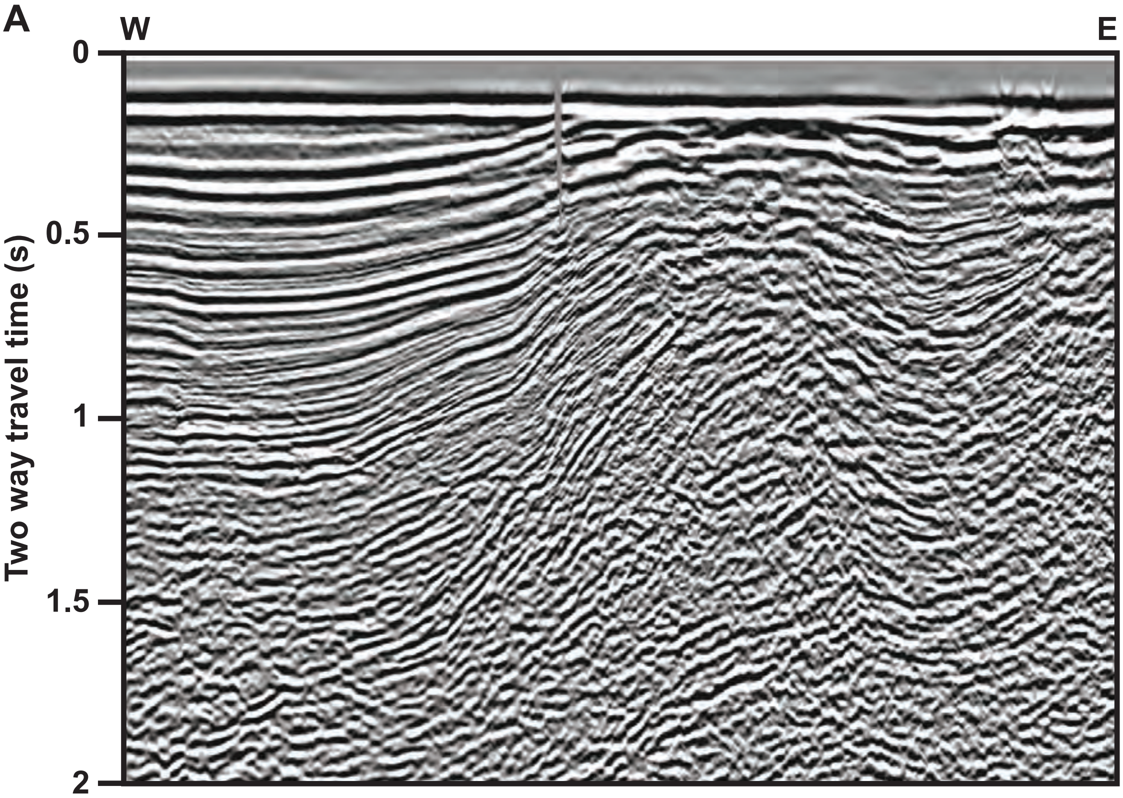
A

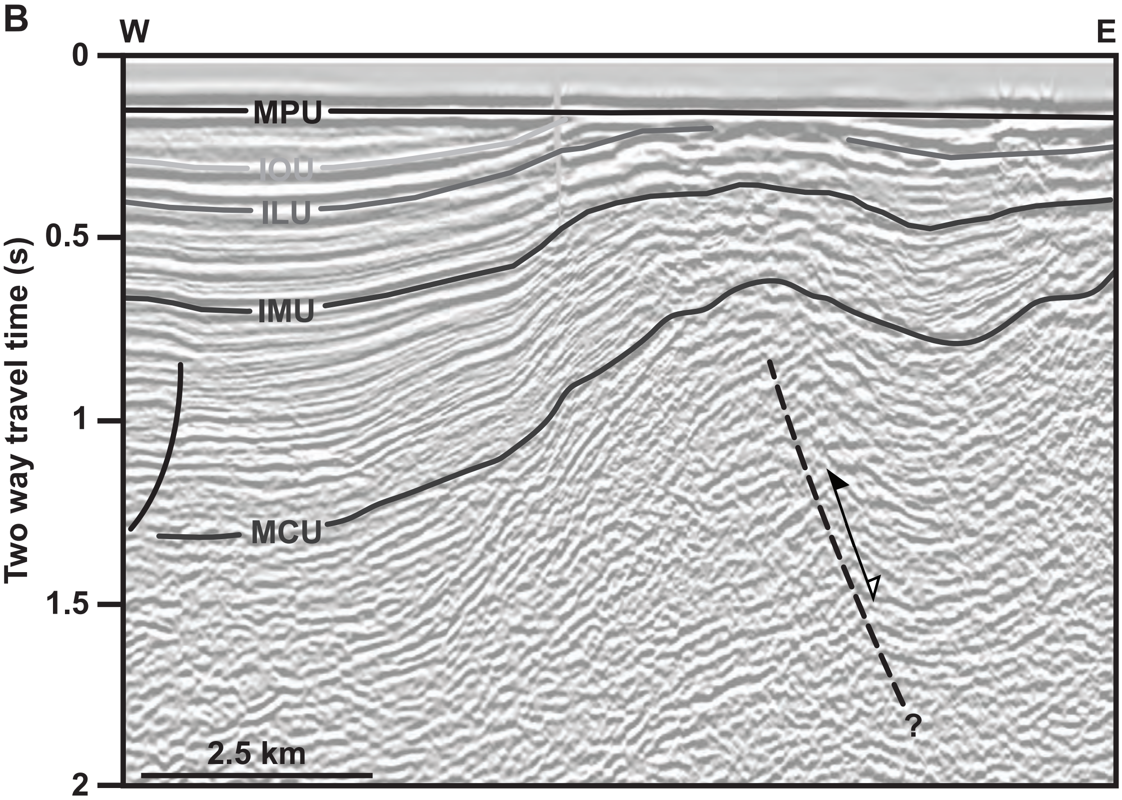


B

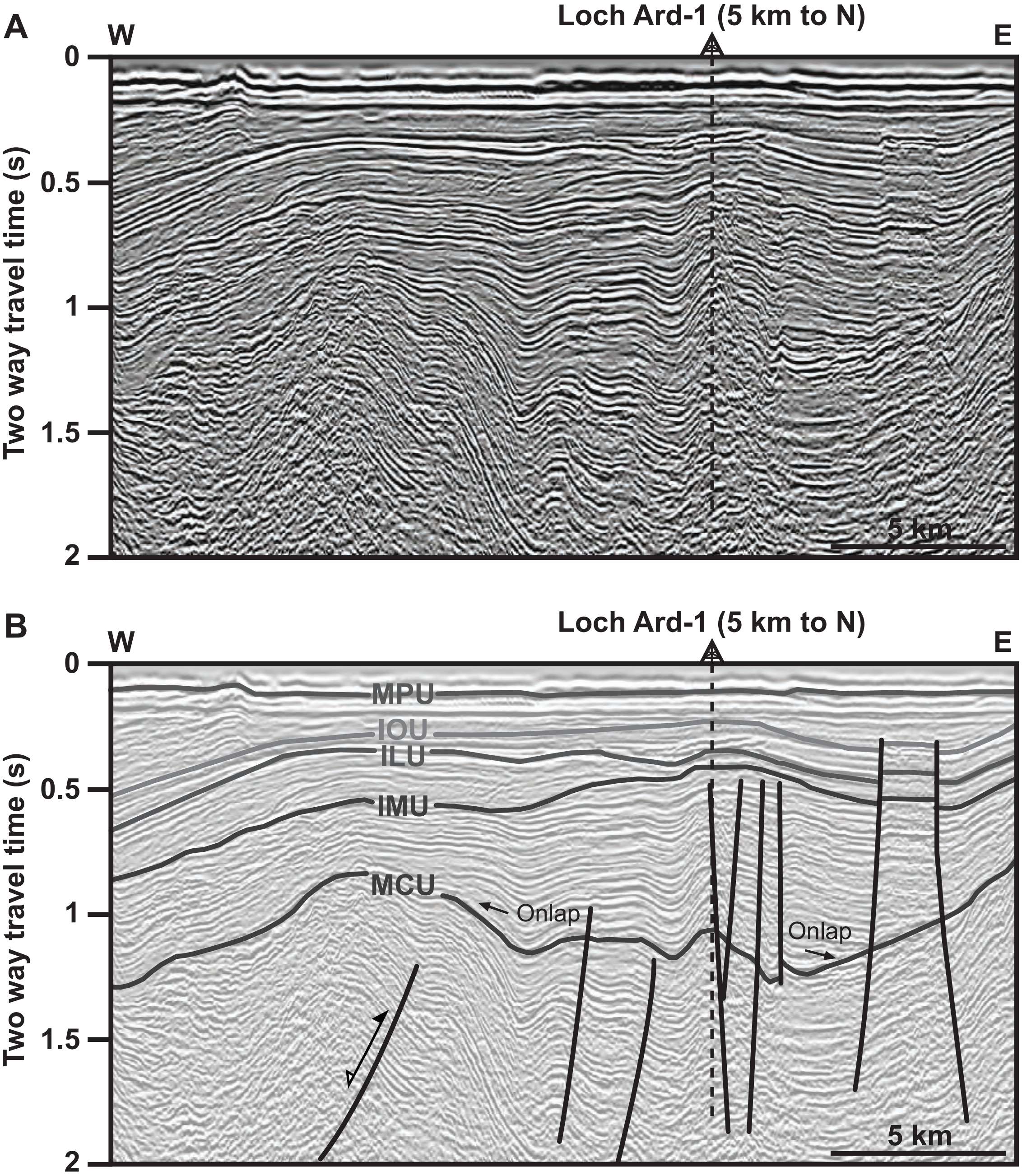


Holford et al Figure 13

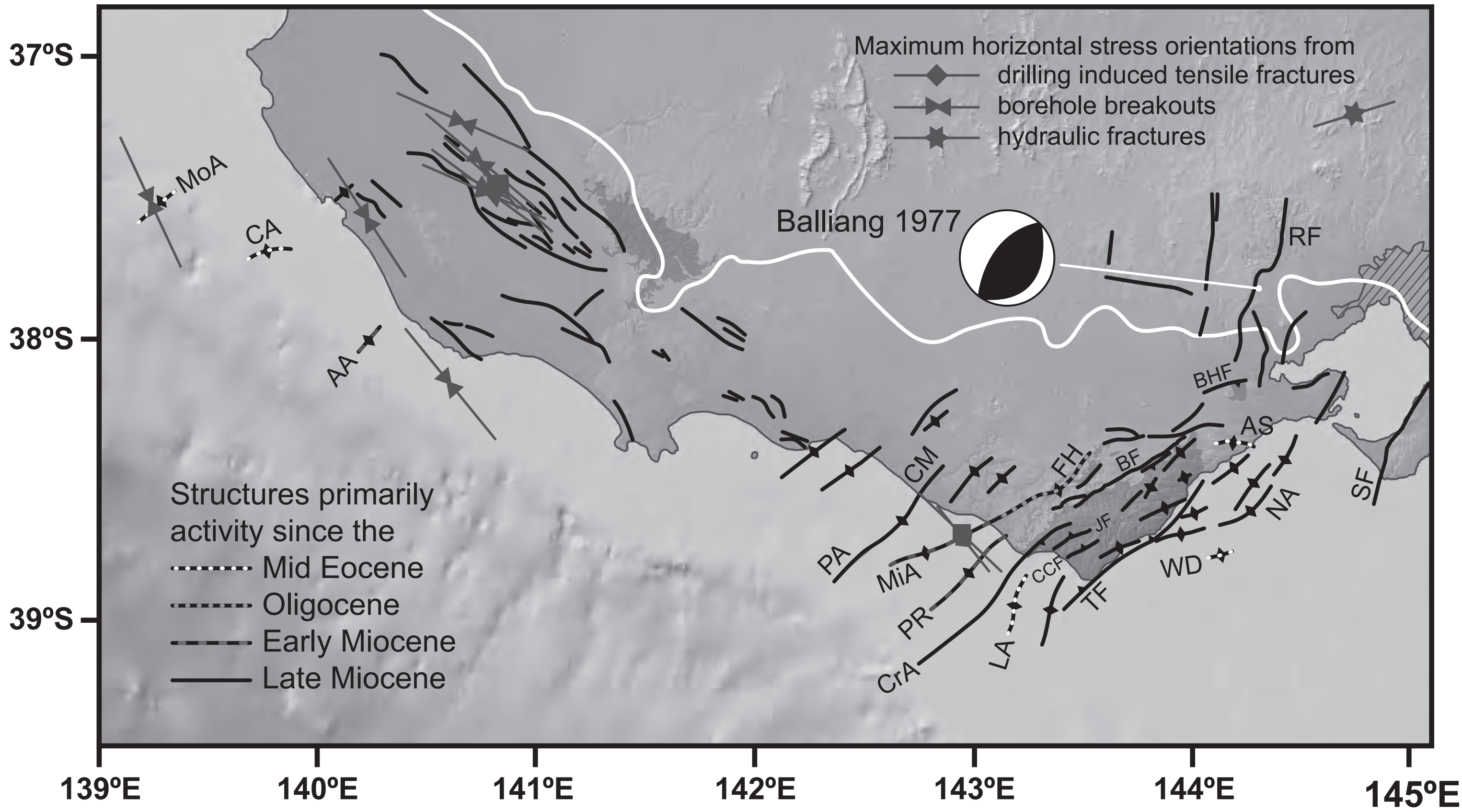




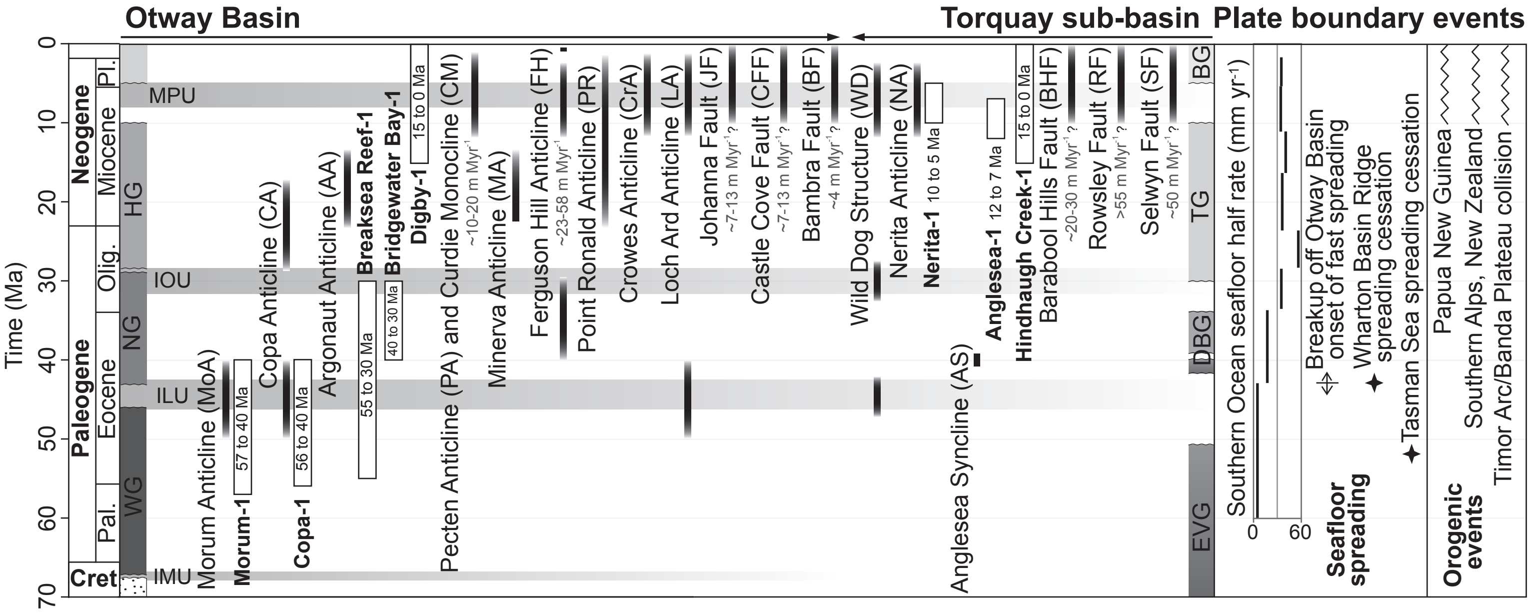
Holford et al Figure 14



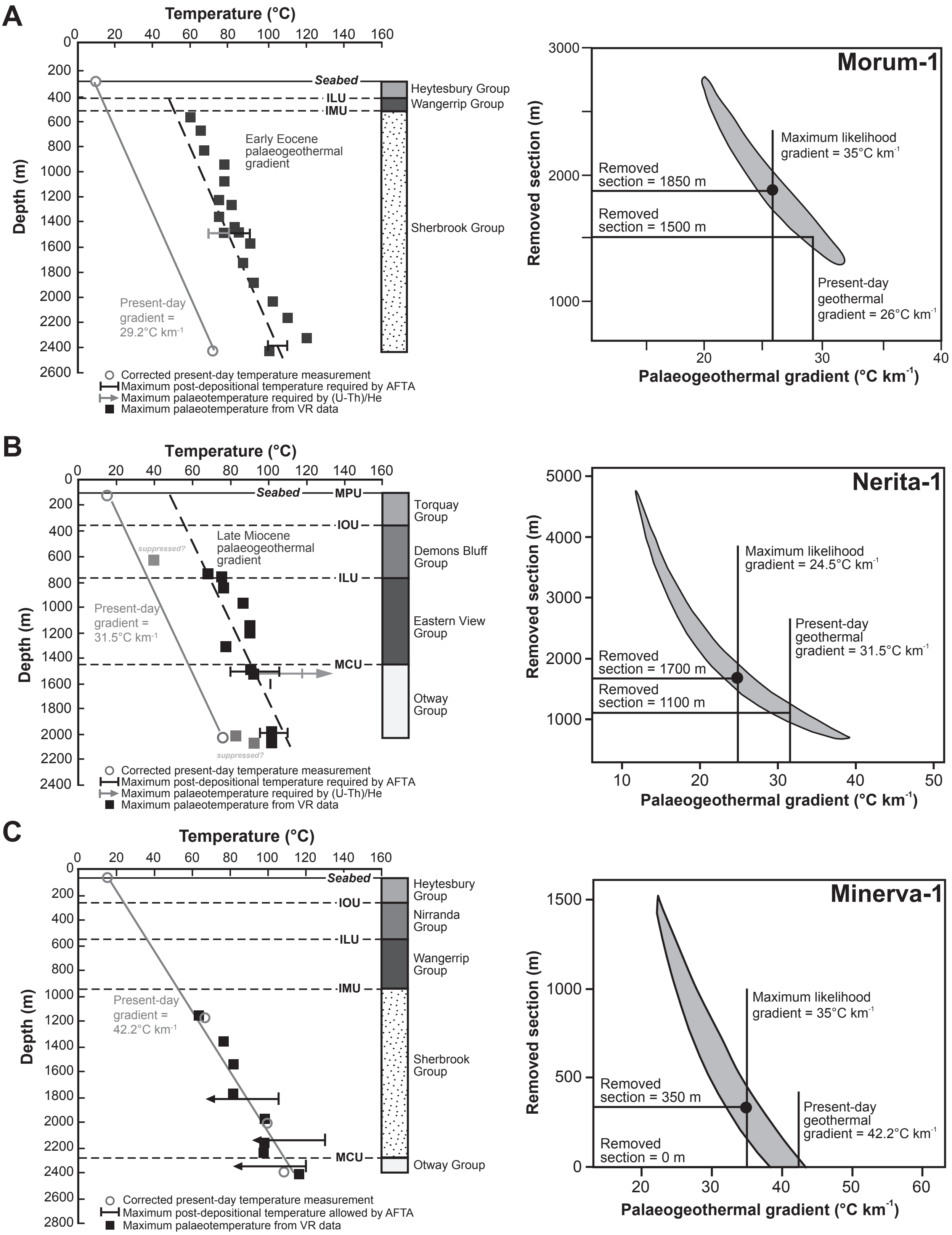
Holford et al Figure 15



Holford et al Figure 16



Holford et al Figure 17



Holford et al Figure 18