Alluvial fans and fan deltas in the Paparoa Formation, Greymouth Basin: a new rift model

Mrinmoy K. Maitra and Kari N. Bassett
School of Earth and Environment, University of Canterbury, Christchurch, New Zealand

ABSTRACT
Sedimentary facies analysis of conglomerate deposits in the Greymouth Rift Basin has identified the latest Cretaceous to Paleocene alluvial fan and fan delta environments on the northwestern side. The Gravelly Braided River Facies Association is interpreted as high energy, braided river streamflow in a streamflow-dominated alluvial fan environment. The Gravelly Delta Front Facies Association was deposited by high bedload mouth bars and channel avulsion. The Gravelly Prodelta Facies Association is interpreted as high-density turbidity currents, and subaqueous debris flows in a fan delta slope environment. Bedding geometries suggest the fan deltas were Hjulström-types formed on lower angle slopes. The gradual decrease in conglomerate thickness from northwest to southeast indicates that the primary basin bounding fault was located offshore to the northwest, most likely the Cape Foulwind-Canoe Fault Zone. Overall facies distribution and paleoflow directions indicate the Greymouth Basin formed as a half-graben in a purely extensional setting with no strike-slip movement. The presence of contemporaneous sub-basins in the West Coast region suggests they likely experienced similar depositional history to the Greymouth Basin. Our findings agree with previous models from the Taranaki Basin that indicate that rifting was purely extensional, suggesting that the West Coast-Taranaki Rift System recorded primarily orthogonal extension.

Introduction
The Greymouth Basin is sub-basin of the West Coast Basin in the West Coast–Taranaki Rift System (80–55 Ma) present in the Taranaki, West Coast, and western Southland regions of New Zealand (Strogen et al. 2017). The Greymouth Basin, filled with the latest Cretaceous to Paleocene Paparoa Formation, has been variably modelled as a sag basin (Gage 1952; Suggate 2014), a complex full-graben rift basin (Bowman 1984), a transtensional rift basin with faults on both sides (Laird 1993, 1994; Bassett et al. 2006), and a half-graben rift basin with faults on both sides (Laird 1993, 1994; Bassett et al. 2006), and a half-graben rift basin with the bounding fault on the eastern side (Newman and Newman 1992; Ward 1997; Kamp et al. 1999). In most cases, the primary syn-depositional fault was postulated as located on the eastern side of the basin because it is the Montgomerie-Mt Davy Fault System marking the eastern boundary (Figure 1; Gage 1952; Newman 1985; Bishop 1992; Ward 1997; Kamp et al. 1999). However, more recent work has suggested that the Montgomerie and Mt Davy faults were not active until after the deposition of the Paparoa Formation and thus could not have been the primary basin bounding fault (Suggate 2014). In addition, the Greymouth Basin contains a thick conglomerate succession on the northwestern side suggesting steeper topography, probably fault controlled (Bassett et al. 2014). This led to our research addressing the questions of (1) where were the primary syn-depositional faults located? and (2) was there a transtensional component to the rifting?

While most previous studies of this basin focused on the characterisation of coal and closely associated rocks (Gage 1952; Bowman 1984; Newman 1985; Newman and Newman 1992; Boyd and Lewis 1995; Ward 1997; Kamp et al. 1999), the thick conglomerate deposits to the northwest, present in both coastal outcrop and drill cores, have received less attention. The large grain sizes (up to boulder) and overall thickness (>400 m) of the conglomerate succession indicate their importance for interpreting the syn-sedimentary tectonics of the basin.

We present here detailed sedimentary facies analysis of the conglomerates and associated finer-grained sediments. Our new research identifies extensive fan deltas alternating with alluvial fans along the northwestern margin of the Greymouth Basin leading to a new interpretation of the paleogeography, location of syn-depositional faults, and geometry of the basin. The results provide new constraints on depositional environments and basin geometry for the wider West Coast-Taranaki Rift System and can serve as...
outcrop facies analogues for other Late Cretaceous rift basins in New Zealand which are deeply buried and only studied via seismic analysis.

**Greymouth Basin**

The Cretaceous break-up of Gondwana was the result of two distinct rift phases (Laird 1981, 1993; Turnbull et al. 1993; King and Thrasher 1996; Gaina et al. 1998; Cook et al. 1999; Laird and Bradshaw 2004; Mortimer 2004; King et al. 2011; Mortimer et al. 2017; Strogen et al. 2017). The older Zealandia Rift phase was during the mid-Cretaceous (c. 105–83 Ma) and produced half-grabens associated with metamorphic core complexes striking NW to WNW; it coincided with the opening of the Southern Ocean and rifting from Antarctica (Figure 2; Strogen et al. 2017). The younger West Coast–Taranaki Rift phase was active during the latest Cretaceous (c. 80–55 Ma) and produced NNE-striking extensional half-grabens; it coincided with active seafloor spreading in the Tasman Sea. These two phases were separated by a short period (c. 83–80 Ma) of uplift and erosion, representing a break-up unconformity.

The sedimentation associated with the West Coast–Taranaki Rift System occurred in northern and southern Taranaki, the West Coast, and the Western Southland regions of New Zealand (King and Thrasher 1996; Hemmings-Sykes 2012; Reilly et al. 2015; Strogen et al. 2017). In the West Coast region, the latest Cretaceous to Paleocene deposition is known as the Paparoa Formation and is found in several localities including Greymouth Coal Field, Pike River Coal Field, Te Kuha (Denniston Plateau), Taku-tai half graben, and Kumara half-graben (Figure 2; Newman 1985; Ferguson 1993; Boyd and Lewis 1995; Ward 1997; Dutton et al. 2013). The Greymouth Sub-basin, the subject of this study, contains more than 800 metres of the Paparoa Formation and is found in several localities including Greymouth Coal Field, Pike River Coal Field, Te Kuha (Denniston Plateau), Taku-tai half graben, and Kumara half-graben (Figure 2; Newman 1985; Ferguson 1993; Boyd and Lewis 1995; Ward 1997; Dutton et al. 2013). The Greymouth Sub-basin, the subject of this study, contains more than 800 metres of the Paparoa Formation and is found in several localities including Greymouth Coal Field, Pike River Coal Field, Te Kuha (Denniston Plateau), Taku-tai half graben, and Kumara half-graben (Figure 2; Newman 1985; Ferguson 1993; Boyd and Lewis 1995; Ward 1997; Dutton et al. 2013). The Greymouth Sub-basin, the subject of this study, contains more than 800 metres of the Paparoa Formation and is found in several localities including Greymouth Coal Field, Pike River Coal Field, Te Kuha (Denniston Plateau), Taku-tai half graben, and Kumara half-graben (Figure 2; Newman 1985; Ferguson 1993; Boyd and Lewis 1995; Ward 1997; Dutton et al. 2013). The Greymouth Sub-basin, the subject of this study, contains more than 800 metres of the Paparoa Formation and is found in several localities including Greymouth Coal Field, Pike River Coal Field, Te Kuha (Denniston Plateau), Taku-tai half graben, and Kumara half-graben (Figure 2; Newman 1985; Ferguson 1993; Boyd and Lewis 1995; Ward 1997; Dutton et al. 2013). The Greymouth Sub-basin, the subject of this study, contains more than 800 metres of the Paparoa Formation and is found in several localities including Greymouth Coal Field, Pike River Coal Field, Te Kuha (Denniston Plateau), Taku-tai half graben, and Kumara half-graben (Figure 2; Newman 1985; Ferguson 1993; Boyd and Lewis 1995; Ward 1997; Dutton et al. 2013). The Greymouth Sub-basin, the subject of this study, contains more than 800 metres of the Paparoa Formation and is found in several localities including Greymouth Coal Field, Pike River Coal Field, Te Kuha (Denniston Plateau), Taku-tai half graben, and Kumara half-graben (Figure 2; Newman 1985; Ferguson 1993; Boyd and Lewis 1995; Ward 1997; Dutton et al. 2013).
Greymouth Basin have been folded and cross-cut by the younger, oblique-slip Montgomerie-Mt Davy Fault System (Suggate 2014).

Tectonic setting and basin geometry of the Greymouth Sub-basin have been interpreted in various ways. Bowman (1984) modelled the Greymouth Basin as a complex full graben basin with two major persistent parallel depocentres separated by a stable horst where sediment thicknesses were largely controlled by NNE-SSW oriented, deep-seated, syn-depositional faults. Newman (1981, 1985), and Ward (1997), as part of their analyses of the coal facies and palynostratigraphy, interpreted the Greymouth Basin as a half-graben with the eastern margin Montgomerie-Mt Davy Fault System responsible for the overall subsidence. An alternative theory suggested that extension of the Greymouth Basin could have been accommodated as oblique movement on NNE–SSW faults, likely reactivated transform faults from spreading in the Tasman Sea (Bishop 1992; Laird 1994), supported by provenance analysis of the Paparoa Formation (Bassett et al. 2006). More recently, Suggate (2014), harking back to Gage (1952), reinterpreted the Greymouth Basin as a down-warping sag basin. He concluded that the Montgomerie-Mt Davy Fault System post-dates Greymouth Basin deposition,
truncating outcrops rather than being a basin bounding fault. This interpretation was based on a revision of the cross-section through the Brunner Anticline suggesting that the onset of uplift was a result of the future Montgomerie-Mt Davy Fault System before deposition of the Brunner Formation during Eocene time (Figure 1).

The basin was infilled with the latest Cretaceous to Paleocene terrestrial sedimentary succession of the Paparoa Formation (Paparoa Coal Measures) characterised by conglomerates, sandstones, coals, coaly mudstones, and mudstones (Gage 1952; Nathan et al. 1986; Newman and Newman 1992; Boyd and Lewis 1995; Ward 1997; Maitra 2020). These have been divided into seven members of alternating coal-bearing sandstone versus massive mudstone lithologies comprising Jay, Ford, Morgan, Waiomo, Rewanui, Goldlight, and Dunollie. The Paparoa Formation unconformably overlies Greenland Group basement rocks of Early Ordovician age and underlies the transgressive succession of Paleocene-Eocene Brunner Formation to calcareous Island Sandstone, Kaiata Mudstone and Oligocene Cobden Limestone of the Haerenga Supergroup (Figure 2; Nathan et al. 1986, 2002; Mortimer et al. 2014).

The typical lithologies of the alluvial members (Jay, Morgan, Rewanui, Dunollie) include sandstones often in channels with strong basal scour and cross-bedding (Gage 1952; Nathan et al. 1986; Newman and Newman 1992; Ward 1997; Maitra 2020). Adjacent to the channels are crevasse splay sandstones interbedded with mudstones with abundant plant detritus and downward bifurcating roots interpreted as floodplains. Channel sandstones commonly fine upward to carbonaceous mudstones, fining further to high ash coals containing abundant detrital plant material and common rootlets (Maitra 2020; Figure 3). The coal-bearing, sandy alluvial deposits are present across most of the basin until truncated in the east by the Montgomerie-Mt Davy Fault System, eroded in the north, and pinched out to the south (Figure 3). They have been extensively studied and interpreted as meandering rivers with adjacent floodplains with oxbow lakes and mires (Gage 1952; Boyd and Lewis 1995; Ward 1997; Maitra 2020). Raised mire coals are commonly thick, waxy and clean with fragmented leaf fossils, and occur in wide lens-shaped units (Newman and Newman 1992; Sherwood et al. 1992; Boyd and Lewis 1995; Maitra 2020). The meandering river and mire facies interfinger with conglomeratic lithofacies to the northwest, the subject of this manuscript, where they are well exposed in outcrops along the coast and in nearby drill holes (Figure 3).

The typical lithologies of the mudstone members (Ford, Waiomo, Goldlight) include massive mudstones of laminated dark grey mudstones and siltstones with common gastropod and mollusc fossils, indicating a freshwater lacustrine setting (Fossil Record Online Database; Gage 1952; Ward 1997). These grades laterally to laminated mudstones with minor to abundant normally graded sandstones with common load casts interpreted as turbidites in a lacustrine setting (Cody 2015; Maitra 2020). Mudstone members are commonly present in the basin centre and east until truncated against the Montgomerie-Mount Davy Fault System, pinch out to the south, or interfinger with conglomerates to the north-west (Figure 3).

The Paparoa Formation is overlain by Paleocene Brunner Formation in the west, whereas it is unconformably overlain by the Eocene Brunner Formation in the centre and eastern side of the basin (Raine 1984; Ward 1997; Nunweek 2001). The Brunner Formation is found across the entire basin except in the coastal Twelve Mile Beach (TMB) outcrop. There the Island Sandstone of Middle Eocene age, which overlies the Brunner Formation of Early Eocene age across the entire basin, sits unconformably on the Paparoa Formation.

Methods

The conglomerates of the Paparoa Formation are well exposed in ~2 km long cliffs along the coast on the northwest side of the Greymouth Basin but also extend inland (<2 km) where they are exposed in scrufty outcrops and intersected by a large number of drill holes used for coal exploration (Bowman 1984; Bowman and Caffyn 1984; Newman 1987; Duff 1988; Breeze and Browne 1999; Boyd 2002; Zehnder 2011). Detailed sedimentary facies analysis of conglomerate and mudstone lithofacies was conducted on the coastal outcrops along Twelve Mile Beach (TMB) and then applied to drill cores inland (Figure 3A).

Key field techniques include (1) measuring bed thickness and geometry, (2) describing clast size, shape, sorting, and composition, (3) identifying sedimentary structures, (4) measuring bedding strike and dip, (5) measuring paleocurrent directions from cross-bedding and clast imbrication, and (6) identifying trace fossils/plant remains. Beds were correlated across small faults following identifiable bedding planes.

Drill core descriptions have been taken from the original logs downloaded from the New Zealand Petroleum and Minerals (NZPM) website (www.nzpam.gov.nz). These have been verified and modified after inspection of the original cores stored in the Featherston Core Store. Thicknesses have been corrected for dip and duplication by faults. Eleven drill holes from the north-western corner of the basin have been used to
develop the sedimentary facies then correlated in order to get a better understanding of their lateral distribution.

**Sedimentary facies analysis**

The main focus of this paper is facies analysis of the conglomerates and associated fine-grained deposits of the Morgan to Dunollie members outcropping in the northwestern corner of the basin. Mudstone deposits of the Waiomo and the Goldlight members crop out at TMB and Seven Mile Creek Road, respectively, as well as being intersected by a number of drill holes within 5 km from the coast.
Conglomerate and mudstone units have been grouped into four facies associations that allow interpretation of depositional environments (Table 1). FA1 (gravelly braided river), FA2 (gravelly delta front), FA3 (gravelly prodelta), and FA4 (lacustrine) will be presented in detail below. A detailed sedimentary facies analysis of the wider Greymouth Basin was developed by Maitra (2020) and forms the basis of the correlations in Figure 3. Summaries of the facies are supplied in the supplementary data file.

Facies association 1 gravelly braided river (FA1)

Description – Facies 1a, conglomerate with minor sandstone, comprises imbricated, clast-supported, rounded to subangular, moderately to poorly sorted, pebble to cobble conglomerate ranging from pebble to boulder (90%) with thin sandstone lenses (10%) (Figure 4A). Beds are 0.6–10 m thick and often occur in large channel lenses with erosional bases. Intra-clasts of laminated, deformed grey to dark grey mudstone and siltstone are present at the base of one particularly large channel (Figure 4B), sourced from the underlying mudstones (Cody 2015). Moderately sorted, medium to coarse sandstones show channel shape geometries and occur in lenses 1–2 metres wide and up to 20 cm thick (Figure 4A). The facies is thick and multi-storeyed. Facies 1b, matrix supported angular conglomerate, is characterised by matrix-supported, angular to subangular, poorly sorted, pebble to cobble conglomerates in muddy to coarse sandy matrix that generally lacks erosional bases. Beds are 0.6–10 m thick. This facies is mostly found in drill holes and is commonly interbedded with facies 1a. Facies 1c, carbonaceous mudstones and sandstones, consists of slightly carbonaceous, silty mudstones to coaly mudstones and interbedded fine sandstones with rare pebbles and low bioturbation (BI:2). There are abundant coaly roots that are vertical and downward branching (Figure 4C). The facies can be found as 60 cm to 1 m thick layers interbedded with facies 1a.

Interpretation – Facies 1a is interpreted to have been deposited in a braided river environment where the dominant process was bedload transport by high energy stream flow (e.g. Kazanci 1988; Collinson 1996; Miall 1996; Miall 2010). Clast imbrication shows the palaeoflow direction towards southeast (Figure 5). Sharp and erosional bases are evidence of channelised flow. The presence of a deeply erosional contact with mudstone intra-clasts at its base indicates down-cutting of a large channel into the underlying lacustrine mudstones. The majority of conglomerates likely were deposited as longitudinal low-relief gravel bars with sandstones deposited in bar-top channels from waning flow (e.g. Todd 1989; Maizels 1993). The subangular clasts and indistinct stratification suggest the clasts travelled a relatively short distance before deposition (e.g. Kazanci 1988; Turkman et al. 2007). Facies 1b is interpreted as deposited by subaerial debris flows (e.g. Nemec and Steel 1984; Tanaka and Maejima 1995; Blair 1999). The mixture of angular to subangular clasts of different grain sizes and no grading suggests deposition may have been proximal to steeper slopes (e.g. Maejima 1988; Tanaka and Maejima 1995). Facies 1c is interpreted to have been deposited in a low-lying overbank to floodplain environment (e.g. Reading 1996) with abundant plant growth based on the common preserved roots. The facies association FA1 represents subaerial deposition in a braided river and floodplain.

Facies association 2 gravelly delta front (FA2)

Description – facies 2a, coarsening upward conglomerate with fitted clasts, mainly consists of imbricated, subangular to subrounded, moderately sorted, conglomerate beds that coarsen and thicken upward from pebble to cobble (Figure 6A). Large cross-beds infill scoured channels with foreset dips that shallow upward into parallel beds (Figure 6A). Cast imbrication is common (Figure 6B). Little to no matrix is present with many clasts showing jigsaw-like fitted edges with sharp contacts and angular indentations on clast interfaces (Figure 6C). Fitted clasts occur in lenses 5–7 m wide and 1–2 m thick. Facies 2b, bioturbated sandstone and mudstone, consists of carbonaceous, highly bioturbated (BI:4), fine-grained sandstones (70%) thinly interbedded at mm to cm scale with highly carbonaceous siltstones (10%) and mudstones (20%). Plant material is scattered or in small (<1 cm) lenses but rootlets are absent (Figure 6C). Carbonaceous siltstones and muddy high ash coal beds are occasionally present. This facies is found interbedded with facies 3a in layers 30 cm to 1 m thick. It is distinguished from facies 1c by the absence of roots, thinner bedding, and high degree of bioturbation.

Interpretation – Facies 2a is interpreted as deposited at the mouth of a high bed load, distributary channel entering a body of water forming a coarsening and thickening upward mouthbar succession (e.g. Reading 1996; Bhattacharya 2010). The basal scours suggest channel incision during high-velocity flow followed by gradual infilling, first by downstream accretion foresets grading to parallel beds (e.g. Galloway and Hobday 1996; Gobo et al. 2014). Large cross-beds show palaeoflow towards southeast (Figure 5). The fitted clast textures in channel lenses may have been formed by microvibrations and abrasion from clast collisions and winnowing of fines from the streamflow in the channels, an interpretation based on work by Schumm and Stevens (1973) and Brewer et al. (1992), although post-burial diagenetic alteration can’t be entirely ruled out (e.g. Vollbrecht et al.
**Table 1.** Conglomerate and associated fine-grained lithofacies.

| Facies Association | Facies name and number | Facies description | Bedding thickness | Depositional processes | Depositional environments | Spatial distribution |
|--------------------|------------------------|--------------------|-------------------|-----------------------|--------------------------|---------------------|
| FA 1 Gravelly braided river | Conglomerate with minor sandstone (1a) | Clast-supported, moderately sorted, subangular to sub-rounded, cobble conglomerates and minor sandstone lenses; shows clast imbrication, erosional channels. | ~1–10 m thick | Bedload transport in a high energy stream flow. | Braided river environment (e.g. Kazanci 1988; Miall 1996; Collinson 1996; Miall 2010) | Northwest |
| | Matrix supported angular conglomerate (1b) | Matrix supported, angular to subangular, poorly sorted, pebble to cobble conglomerates in muddy matrix; lack erosional bases. | 35–60 cm | Subaqueous debris flow | Subaqueous alluvial fan and fan delta plain proximal to steep slopes (e.g. Maejima 1988; Tanaka and Maejima 1995) | |
| | Carbonaceous mudstones and sandstones (1c) | Carbonaceous silty mudstones and fine sandstones, rare pebbles, low bioturbation, pyrite and roots. | 60 cm to 1 m | Suspension settling; river flooding into low-lying areas such as abandoned channels | Overbank to floodplain environment (e.g. Reading 1996) | |
| FA 2 Gravelly delta front | Conglomerate with foresets and fitted clasts (2a) | Coarsening upward, clast supported, pebble to cobble conglomerates; channels filled with cross-beds that shallow in dips upwards; shows fitted clast textures. | 1–3 m thick | High bed load, gravelly, prograding mouthbars, channel incision with downstream accretion, jostling and abrasion between clasts. | Mouthbar in a gravelly river channel (e.g. Reading 1996; Bhattacharya 2010) | Northwest |
| | Bioturbated sandstone and mudstone (2b) | Carbonaceous, highly bioturbated fine to medium-grained sandstones and mudstones with scattered plant material. | 30 cm to 1 m | Mainly from suspension, burrows | Subaqueous, low energy interdistributary bay environment favoured by burrowing organisms (e.g. Jones and Schumm 1999). | |
| FA 3 Gravelly prodelta | Conglomerate with deformed sandstone lenses (3a) | Clast supported, pebble to cobble conglomerates with sandstone lenses, load casts, and abundant soft-sediment deformation. | 2–3 m | High-density bedload transport in a gravel dominated river discharge, sediment gravity flows from slope failure, and fluidisation. | Upper fan delta slope (e.g. Anadón et al. 1991; Renaut and Gierlowski-Kordesch 2010). | Northwest |
| | Matrix supported rounded conglomerate (3b) | Matrix supported conglomerate to pebbly mudstone. Clasts are pebble to cobble sized, subrounded to well-rounded. Matrix is carbonaceous mudstone. | 10–25 cm | | | |
| | Mudstone with normally graded sandstone and conglomerate lenses (3c) | Mudstones and siltstones interbedded with normally graded sandstone beds, conglomerate lenses, and dropstones. | | | | |
| FA 4 Lacustrine | Massive mudstone (4a) | Massive, highly carbonaceous mudstones with freshwater molluscs from *Hyridella* species (Gage 1952) and leaves of genus *Banksiaformis* (Hill and Christophel 1988) | Mudstone beds are mm to cm thick lamina collectively in thicker layers (mostly 10–12 m thick) | Suspension settling | Freshwater lake (Gage 1952; Ward 1997; Cody 2015; Maitra 2020; e.g. Anadón et al. 1991; Renaut and Gierlowski-Kordesch 2010; FRED 2015) | Basin centre, northeast and southeast |
| | Mudstone with minor thin sandstone (4b) | Carbonaceous mudstones (90%+) interbedded with thin, normally graded, very | Mudstone beds are mm thick lamina collectively in thicker layers | Suspension settling with irregular and rare turbidity currents | | |
The association of the interlocking clasts with the coarsening upward beds in channel lenses indicates that the jostling and abrasion took place at the mouth of a channel. Facies 2b is interpreted to have been deposited in a subaqueous, low energy environment with relatively high bioturbation by burrowing organisms (e.g. Jones and Hajek 2007), likely in interdistributary bays. The closely interbedded nature of Facies 2a and 2b indicates channel avulsion was common (e.g. Jones and Schumm 1999). We therefore interpret FA2 as deposited in a gravelly delta front with prograding and avulsing channel mouthbars and interdistributary bays (e.g. Jones and Schumm 1999; Mohrig et al. 2000; Field 2001; Swenson 2005).

**Facies association 3 gravelly prodelta (FA3)**

*Description – facies 3a, conglomerate with deformed sandstone beds, consists primarily of clast supported,
subangular to subrounded, moderately sorted, pebble to cobble conglomerates (85%) interbedded with minor fine to medium sandstone lenses (15%) commonly showing soft-sediment deformation (Figure 7A). The facies occurs in thick, multi-storeyed packages with beds 3–7 m thick. Inversely graded beds and pebbles lining the foresets of sandy crossbeds are rare but present. The abundant soft-sediment deformation includes load casts, convolute bedding, flame structures, and fluid escape features in both

Figure 6. FA2 gravelly delta front at Twelve Mile Beach; A, gravelly mouthbar channel with infilling crossbeds at the base and planar beds at the top in facies 2a, B, clast imbrication in gravelly mouthbar channel showing paleocurrent orientation towards southeast in facies 2a which is underlain by carbonaceous, bioturbated sandstone and mudstone of facies 2b, C, angular indentations on clast interfaces showing fitted clast texture in a mouthbar channel in facies 2a.
the conglomerate and sandstone beds. Facies 2a is easily identified in outcrop by the large soft-sediment deformation but is, however, difficult to distinguish in cores due to their small diameter. Facies 3b, matrix-supported rounded conglomerate, comprises pebble to cobble sized, subrounded to well-rounded conglomerate clasts (30%) in a carbonaceous muddy matrix (70%) (Figure 7B). Beds are relatively thin ranging from 10 to 25 cm. This facies is rare but present both in outcrop and drill cores where it is easily identifiable and commonly associated with facies 3a. Facies 3c, mudstone with normally graded sandstone and conglomerate, comprises silty mudstone (60%) interbedded with sandstones (30%) and conglomerate lenses (10%) (Figure 7C). Mudstone layers are laminated, and range from 20 cm to 1 m thick. Sandstone beds are commonly 20–30 cm thick, often normally graded, sometimes with a basal pebble lag or coarse sandstone base, but more commonly fining from medium to fine sandstone to siltstone. Isolated pebbles or cobbles occur singly either at the bases of sandstone beds or indenting layers in the laminated siltstones (Figure 7D). Conglomerates are mostly pebble in size, subangular to subrounded, clast-supported, normally graded, and occur in lenses 30–90 cm thick and 1–5 m wide, commonly with erosional basal scours. This facies is easy to distinguish in both outcrop and core as indicated by the presence of normally graded sandstones and conglomerate lenses.

Interpretation – Facies 3a is interpreted as being deposited on a slope in a subaqueous environment based on the abundant load casts and other soft-sediment deformation. The dominance of pebble to cobble conglomerates suggests high-bedload stream flow entering a body of water subsequently remobilised by sediment gravity flows from slope failure (e.g. Lowe 1975; Allen 1982; Mills 1983; Van Loon and Brodzikowski 1987; Van Loon 1992; Owen 1996; Stromberg and Bluck 1998; Maltman and Bolton 2003; Owen and Moretti 2011). Pebble-lined cross-beds show the palaeoflow direction towards southeast (Figure 5). Facies 3b is interpreted as subaqueous debris flows formed by mobilisation of loose sediments in a delta slope region based on matrix supported yet well-rounded clasts in a carbonaceous muddy matrix (e.g. Nemec and Steel 1984; Blair 1999; Miall 2010). Facies 3c is interpreted to have been deposited by turbidity currents ranging from sand to pebble dominated based on their interbedded, normally graded nature (e.g. Makaske et al. 2002; Boggs Jr 2006; Jones and Hajek 2007; Renaut and Gierlowski-Kordesch 2010). Isolated conglomerate clasts that indent underlying layers are interpreted as dropstones (e.g. Gilbert 1990; Bennett et al. 1996) most likely
transported by vegetation rafting (e.g. Postma et al. 1988) or floating logs (e.g. Emery 1955, 1965; Bennett et al. 1996) rather than icebergs since recent leaf fossil and beech tree pollen studies indicate the climate was too warm to support major bodies of ice (Ward 1997; Kennedy 2003; Raine et al. 2017). Taken together, we interpret a gravelly prodelta environment in a gravelly delta setting (e.g. Nemec and Steel 1984; Postma 1984, 1990; Prior et al. 1986; Martinsen 1989; Nemec 1990; Stow et al. 1996; Boggs Jr 2006; Hilbe and Anselmetti 2014).

Facies association 4 lacustrine (FA4)

Description – facies 4a, massive mudstone, consists of highly carbonaceous, grey to black mudstone/silty mudstone with occasional preservation of pyrites and tuff beds (Figure 8A). The facies occurs in thick deposits ranging from minimum 15 m to maximum 180 m. Small leaves, pollen and freshwater fossils may be locally abundant including freshwater gastropods from Hyridella species, Banksiaformis leaves, and pollen from Nothofagus (southern beech) species (Figure 8B and C; Gage 1952; Ward 1997; Kennedy 2003). Recent microscopic analysis reveals the presence of Botryococcus algae in mudstones (Mohnhoff et al. 2017). Facies 4b, mudstone with minor thin sandstone is characterised by carbonaceous mudstones interbedded with thin, very fine sandstone beds (10% sandstone, 90% mudstone) (Figure 8D). Sandstone beds look similar to those of the lacustrine massive mudstone facies whereas sandstone beds are very fine, normally graded with occasional ripple cross-laminations and are less than 3 cm thick. Basal contacts of the thin sandstone are commonly erosional or sometimes wavy and sharp whereas the upper contacts are slightly gradational with overlying mudstones. Conspicuous plant debris (~10%) is common with rare leaves of genus Banksiaformis (Hill and Christophel 1988) present in sandstone beds. Both facies are widely distributed throughout the Greymouth Basin and are easily identified both in outcrop and drill core.

Interpretation – facies 4a is interpreted to have been deposited in a fresh water lake by relatively slow sedimentation from suspension beyond the direct influence of shoreline or deltaic processes as indicated by the presence of Hyridella species molluscan fossils.

Figure 8. FA4 Lacustrine facies association: A, massive mudstone facies 4a of the Goldlight Member at Spring Creek Mine (Seven Mile Stream), B, leaf fossils at 289.65 m in DH-624, C, fresh water mollusc fossils at 293.98 m in DH-624, and D, mudstone with minor thin sandstone facies 4b of the Waiomo Member at Twelve Mile Beach.
and the absence of sandstone beds (Gage 1952; Ward 1997; Cody 2015; FRED 2015; Maitra 2020). The presence of Botryococcus algae indicates that sedimentation occurred in a deep temperate or tropical lake as the algae often floats in large masses and blooms in the presence of elevated levels of dissolved inorganic phosphorus (e.g. Wolf et al. 1985; Metzger and Largeau 2005). Facies 4b is interpreted to have been deposited in a lake distal from the shoreline but within the influence of turbidity currents as indicated by the presence of normally graded, very fine-grained sandstone beds and the abundance of dispersed plant debris and displaced gastropods and bivalves (Fossil Record Online Database; Gage 1952; Ward 1997; Maitra 2020). The presence of pyrite indicates an anoxic environment (Marnette et al. 1993; Suits and Wilkin 1998; Schoonen 2004; Wang et al. 2013). The presence of leaf fossils of genus Banksiaeformis and pollen such as Nothofagus suggests that the Late Cretaceous climate was quite similar to that of today (Hill and Christophel 1988; Ward 1997; Kennedy 2003; Raine et al. 2017).

**Stratigraphic and lateral distribution of sedimentary facies**

The best exposure of all conglomerate facies is along Twelve Mile Beach (TMB) where there is continuous section on the shore platform and cliff face of the Morgan to Dunollie members of the Paparoa Formation (Figure 9). Outcrop correlates to multiple drill cores along the northwestern side of the basin (Figure 10). The overlying stratigraphy, visible in the shore platform between the Morgan Member and underlying Greenland Group (Figure 10). The basal contact at TMB is an angular unconformity visible in the shore platform between the Morgan Member and underlying Greenland Group basement rocks (Figure 9). Inland in the north, the drillholes that penetrate to basement all contain Morgan Member conglomerates. Several don’t penetrate to basement and so provide no information. Interestingly, further south, the drillholes penetrate basement but the younger Rewanui Member sits unconformably on Greenland Group (Figure 10).

The overlying stratigraphy, visible in the shore platform at TMB, is assigned to the Morgan Member and is interpreted as FA1 gravelly braided river as indicated by the presence of interbedded facies 1a with rare 1b and 1c. This grades upward into FA2 gravelly delta front starting with coarsening upward successions of alternating facies 2a and 2b. A muddy, high ash, coal layer 30 cm thick suggests deposition probably took place in a subaerial floodplain environment near the delta front (at 29.7 m of Figure 9). Nearby DH-619, south of TMB, also contains FA1 (Figure 10). Further inland, FA1 and FA2 correlate laterally with coal-bearing meandering river facies of the Morgan Member toward the basin centre (Figure 3).

There is a gradual contact between the alluvial Morgan Member and the overlying lacustrine Waiomo Member, seen at TMB as multiple coarsening upward successions of alternating facies 2b to 2a (FA2 gravelly delta front), grading to fining upward successions of facies 3c (FA3 gravelly prodelta), eventually fining to ~15 m of facies 4b then 4a (FA4 lacustrine) (Figure 9). The lacustrine mudstones are overlain by FA3 gravelly prodelta grading up to FA2 gravelly delta front although still assigned to the Waiomo Member, indicating progradation of a gravelly delta. The lacustrine mudstone FA4 and gravelly prodelta FA3 correlate laterally to all cores that penetrate to the underlying Morgan Member (Figure 10) providing a well-established marker bed for correlation.

At TMB, the contact with the overlying Rewanui Member is sharp and deeply erosional, marked by a ~1 m thick layer of mudstone intra-clasts from the underlying Waiomo Member (Figures 4B and 9). The overlying ~10 m of conglomerate comprises primarily facies 1a of FA1 gravelly braided river with a prominent layer of facies 1c with abundant roots preserved. FA1, mostly facies 1a with occasional 1b and 1c, correlates to almost every drill core with thickness decreasing from northwest to southeast (Figure 10). Further inland, FA1 correlates with the coal-bearing, sandy meandering river facies of the productive Rewanui Member (Figure 3).

The subaerial FA1 gravelly braided river environment at TMB grades upward into FA2 gravelly delta front (mostly 2a and 2b) indicating the presence of a standing body of water to the east, in this case a lake, recorded by the lacustrine massive mudstones of the Goldlight Member that are present across most of the Greymouth Basin (Figure 3). For the next ~290 m, FA2 gravelly delta front alternates with FA3 gravelly prodelta with only ~10 m of FA1 gravelly braided river present from ~112.5 to 120 m (Figure 9). Facies 2a and 3a dominate the succession suggesting channel avulsion in the gravelly delta and/or slight changes in lake level. This part of the conglomerate succession at TMB is included in the Rewanui Member but we consider it to be correlative to the Goldlight Member and have labelled it so. Inland, FA2 gravelly delta front can be identified in the drill holes by the coarsening upward successions of alternating facies
2a and 2b (Figure 10). FA3 gravelly prodelta is more difficult to recognise due to core diameter being too small to recognise the load casts which are the defining features of facies 3a leading to these possibly being misidentified. The lacustrine mudstones and shallower water normally graded sandstones of the Goldlight Member are found in most cores except those nearest TMB. The ∼110 m thickness of

Figure 9. Twelve Mile Beach stratigraphic column. Depths are in metres with 50 cm divisions.
mudstones in DH-636 to the southeast correlates with the ∼290 m thickness of conglomerates and sandstones at TMB.

The uppermost 25 m of conglomerates at TMB are assigned to the Dunollie Member and are interpreted as FA1 gravely braided river (mostly 1a). FA1 outcrop is bleached nearly white yet is the same composition and texture as all underlying conglomerates. Inland in core, FA1 is dominated by facies 1a with occasional interbeds of facies 1b and 1c. To the east and south, the most abundant facies are the correlative coal-bearing, sandy meandering river facies (Figure 10). This change marks the transition from a lacustrine setting in the centre of the basin to meandering rivers and mires (Figure 3).

At TMB, the Dunollie Member is unconformably overlain by the glauconitic, fossiliferous, calcareous Island Sandstone of Middle Eocene age (Raine 1984; Newman 1985; Nunweek 2001; Monteith 2015). The unconformity represents the onset of the marine
transgression and was mostly likely formed by wave erosion on the transgressive surface. We interpret the underlying highly bleached section as a weathered surface below a thick coastal coal swamp. Unfortunately, the unconformity removed the overlying coals so this is somewhat speculative although it matches nearby stratigraphy of a deep weathering profile beneath coals (Newman 1985; Monteith 2015). In the basin centre, the Dunollie Member is gradationally overlain by the coals, coarse sandstones, and conglomerates of the Brunner Formation of Palaeocene to Early Eocene age which is unconformably overlain by the Eocene Island Sandstone (Raine 1984; Newman 1985; Nunweek 2001).
Interpretation of sedimentary environments

Alluvial fans

FA1 gravelly braided river was likely deposited by steamflow dominated alluvial fans created by steeper slopes to the northwest, most likely controlled by faults (e.g. McPherson et al. 1988; Horton and DeCelles 2001; Blair 2002; Blair and McPherson 2009; Weissmann et al. 2010). Facies 1a and 1b are typical of simple braided river deposition with the dominantly cobble size and sub-rounded clasts suggesting moderate transport distances. However, the presence of boulder class sizes in facies 1a in the lower Rewanui Member and angular to subangular clasts in the debris flow facies 1b in both Morgan and Dunollie members, suggest a relatively shorter distance to the steeper slopes. The steamflow dominated alluvial fans were correlative with the coal-bearing sandy meandering rivers and floodplains located toward the southeast in the alluvial Morgan, Rewanui and Dunollie members.

Fan deltas

FA2 gravely delta front and FA3 gravely prodelta indicate the steamflow-dominated alluvial fans intermittently prograded into a lake forming fan deltas during deposition of the Waiomo and Goldlight members (e.g. Nemec and Steel 1984; McPherson et al. 1988; Harvey et al. 2005; Rees et al. 2018). The internal structure and facies distribution lacks clear distinction between topsets, foresets and bottomsets (e.g. Postma 1990, 2003) and the delta slope lacks steeply dipping foresets such as found in Gilbert-style fan deltas (e.g. Galloway and Hobday 1996; Gobo et al. 2014). Therefore, we interpret the fan deltas as Hjulström-type fan deltas formed on a relatively low angle slope with constantly flowing streams (e.g. Keresztfi et al. 2015). The delta front and subaqueous prodelta probably had a similar gradient to its delta plain.

The existence of gravelly fan deltas during both Waiomo and Goldlight member deposition indicates different times when lakes were present in the basin and the stream-dominated alluvial fans changed to Hjulström-type fan deltas. The relative thickness of fan delta deposits correlate with the Goldlight Member at TMB is about 12 times greater than those of the Waiomo Member (Figures 9). This suggests that fan deltas in the Goldlight Lake probably existed for either a significantly longer period of time or accommodation increased. Unfortunately, the dating resolution isn’t high enough to distinguish the two options (Ward 1997) although the greater thickness of correlative lacustrine mudstone in the Goldlight Member suggests a longer time rather than greater accommodation.

Interpretation of tectonic setting

Northwest location of basin-bounding fault

Thickening of alluvial fan and fan delta facies to the northwest (Figures 3, 9 and 10) and south-eastward paleoflow measured from clast imbrication and cross-bedding in the conglomerate facies (Figure 5) suggest source areas and steeper topography were to the northwest, most likely from a south-east dipping normal fault. Given the lack of coarser sediment supply from the south-eastern side of the basin (Figure 3), we interpret the north-western fault as the major basin-bounding fault in the Greymouth Rift Basin for the Morgan to Dunollie members.

The most probable location of the basin bounding fault was offshore. One of the petroleum wells (Haku-1), located ~15 km northwest of TMB, comprises very thin Eocene Brunner Formation unconformably overlying Greenland Group basement rocks (Figures 1 and 2; Hematite Petroleum 1970) indicating the main basin boundary fault must have been located between the Haku-1 well and TMB. The most likely candidate is the Cape Foulwind-Canoe Fault Zone, mapped as an active marine fault and interpreted as an NNE-SSW trending normal fault reactivated as a reverse fault in the modern compressional tectonic setting (Figure 1; Barnes and Ghisetti 2013).

Half-Graben tectonic model

The distribution of alluvial fan/fan delta facies along the north-western side of the basin interfingering with coal-bearing meandering rivers or lakes in the basin centre and the existence of sandy meandering rivers on a low gradient topographic slope to the south-southeast (Figure 3; Maitra 2020) are typical of a classic half-graben geometry (e.g. Leeder and Gawthorpe 1987; Gawthorpe and Leeder 2000). Paleocurrent measurements from basin centre coal-bearing sandstone facies (cross-bedding and channel axis orientations) show south-westward flowing meandering rivers (Figure 5). The paleoflow direction was parallel to the fault controlled, steep topography indicating axial flow. The presence of the coal mires in the basin centre over- and under-lying lacustrine facies indicates the basin’s depocentre was adjacent to the fault. Periodic motion on the basin bounding fault would have maintained the actively rising highlands to the northwest. The ~400 m thickness of gravelly facies indicates that sediment supply was great enough to fill available accommodation and to keep the lacustrine shoreline in the same location for several million years. A half-graben model has been chosen, as opposed to a full graben model, based on the lack of coarse-grained material on
the eastern side of the basin that would indicate the presence of a second basin-bounding fault. It also better fits the facies distributions where axial meandering river channels and the deepest lake deposits interfinger with the alluvial fans and fan deltas indicating maximum subsidence adjacent to a basin-bounding fault to the north-west.

Our tectonic models for the Greymouth Basin illustrate a half graben where the faulted foot-wall margin was steep and produced high energy flows carrying large volumes of coarse-grained sediments deposited in the alluvial fans and fan deltas along the northwestern side of the basin (Figure 11). Model A shows lakes of the Waipoua and Goldlight members filling the depocentre, interfingerig with fan deltas on the faulted side and with sandy meandering rivers on the low gradient hinge side (Figure 3). Model B is more appropriate for when axial meandering river systems and mires were present in the centre of the basin (Figure 11). Differential subsidence would have caused channel migration through time keeping the main channel adjacent to the alluvial fans (e.g. Leeder and Gawthorpe 1987).

Whether lacustrine or meandering alluvial facies were deposited in the depocentre depended on the balance between the creation of accommodation by subsidence and the amount of sediment supplied to fill it (Leeder and Gawthorpe 1987; Shanley and McCabe 1994; Martins-Neto and Catuneanu 2010; Holz et al. 2015, 2017). The presence of lacustrine settings in the basin depocentre indicates there were phases of high accommodation creation (e.g. Martins-Neto and Catuneanu 2010; Holz et al. 2015, 2017) when accommodation was greater than sediment supply, probably due to periods of more rapid subsidence due to more rapid fault motion. Alluvial deposition in the basin centre indicates sediment supply was greater than or equal to accommodation suggesting decreased subsidence rate from less activity on basin-bounding faults. Cyclic changes in sediment supply, perhaps due to changes in rainfall, are rejected as less likely based on analyses of paleoclimate which show there were no cyclical climatic changes happening in the Late Cretaceous to Paleocene (Ward 1997; Kennedy 2003; Raine et al. 2017).

Seismic facies in the Offshore Takutai Half-Graben

The offshore Takutai Half-Graben is located to the south-west of the Greymouth Basin and formed at the same time so can be used to interpret the broader tectonic setting (Figures 1 and 12). The steeper eastern side of the Takutai Half-Graben is bounded by a major WNW striking normal fault which is truncated by the Cape Foulwind-Canoe Fault Zone (Bishop 1992; Suggate 2013). Reflectors dip and thicken to the SE toward the WNW dipping fault forming fanning bedding (Figure 12; Maitra 2020). The fanning dips become more parallel and horizontally bedded up-section. We have interpreted this as half-graben geometry with the basin bounding, normal fault on the southeastern side, later reactivated as the reverse Cape-Foulwind-Canoe Fault Zone.

Our seismic facies analysis of the Takutai Half-Graben identifies the basin was probably filled with facies similar to those in the Greymouth Basin (Figure 12). We interpret the transparent reflectors in the seismic profile as lacustrine facies, whereas the strong reflectors suggest the presence of coals and alluvial sediments (e.g. Barrier and Nicol 2017). Fault proximal alluvial fans and fan deltas are visible on the right side of the seismic profile and hinge side sandy/muddy river delta/meandering alluvial and floodplain systems are visible on the left side. Both grade laterally to basin centre transparent lacustrine mudstone seismic facies alternating with more reflective meandering alluvial systems and adjacent raised mires seismic facies.

The Cape Foulwind-Canoe Fault Zone was the likely primary basin bounding fault for both the Greymouth Basin and the Takutai Half-Graben indicating the two half-grabens might have been linked by an inter-basin transfer fault (Figure 13; e.g. Gawthorpe and Hurst 1993). Such inter-basin faults link individual half-grabens where major border faults are located on opposite sides of a rift with fault dips in opposite directions (Gibbs 1984; Leeder and Gawthorpe 1987; Rosendahl 1987; Larsen 1988; Peacock and Sanderson 1991; Gawthorpe and Hurst 1993). Examples of these features are found in Lake Tanganyika and Lake Malawi in the East African Rift, in the Red Sea of the Gulf of Suez Rift, and in the Gulf of Evvia in central Greece (Ebinger et al. 1987; Rosendahl 1987; Gawthorpe et al. 1990; Roberts and Jackson 1991).

Discussion

Changed location of primary basin-bounding fault

The evidence justifies that the Greymouth Basin was formed in a fault-controlled basin with the major basin bounding fault located on the north-western side. Most of the previous models depicted the eastern margin Montgomerie-Mt Davy Fault System as the basin bounding fault responsible for overall subsidence (Newman 1985; Ward 1997; Kamp et al. 1999). However, Suggate (2014) showed that the Montgomerie-Mt Davy Fault System resulted from Neogene inversion and was not active until after the deposition of the Eocene Brunner Formation,
and thus could not have been the primary basin bounding fault for the majority of the basin’s subsidence history. Our facies analysis confirms the Montgomerie-Mt Davy Fault System was a younger, cross-cutting fault system not active at the time of deposition (Figure 3). Suggate’s (2014) suggestion that the Greymouth Basin was therefore a sag basin is rejected given the evidence for fault-controlled subsidence and coarse-grained sediment supply from the northwest.

Figure 11. Tectonic models of the Greymouth Basin. Both models illustrate a half-graben geometry with a highland area and basin bounding fault to the northwest forming a steep gradient down to the low-lying basin centre. Model A, illustrates fan deltas entering a basin centre paleolake. Model B illustrates alluvial fans interfingering with meandering river and floodplain depositional systems in the basin centre.
A. Seismic reflection character and facies identification

| Facies                        | Amplitude     | Lateral continuity | Geometry                                                | Seismic Image                                      |
|-------------------------------|---------------|--------------------|---------------------------------------------------------|----------------------------------------------------|
| Alluvial fan/debris flow      | Low-high      | Discontinuous to relatively continuous | Chaotic, steeply dipping, wedge-like                    |                                                    |
| Mouthbar conglomerate         | Moderate to high | Discontinuous to relatively continuous | Channel-shaped, low to moderate dipping                 |                                                    |
| Gravelly turbidite            | Variable      | Disrupted to discontinuous | Chaotic, lens to wedge-like                            |                                                    |
| Mouthbar sandstone            | Moderate to high | Relatively continuous | Channel-shaped, low dipping, sub-parallel              |                                                    |
| Lacustrine mudstone           | Low           | Relatively continuous | Sub-parallel                                           |                                                    |
| Sandy turbidite               | Low           | Disrupted to discontinuous | Chaotic, lens to wedge-like                            |                                                    |
| Meandering river sandstone and floodplain mudstone | Moderate to high | Relatively continuous | Channel fill, sub-parallel                            |                                                    |

B. Seismic profile (P059-84-02) of the Takutai Basin

C. Interpreted seismic profile (P059-84-02) of the Takutai Basin

Figure 12. Offshore seismic image line P059-84-02 (shown in Figure 1B; Bishop 1992; Suggate 2014) is located to the southwest of the Greymouth Sub-basin. A, Seismic facies characteristics and corresponding interpreted alluvial and lacustrine sedimentary facies. B, Uninterpreted seismic profile showing half-graben geometries, interpreted as the Takutai half-graben by Bishop (1992) and Suggate (2013). C, Interpreted seismic profile showing multiple normal faults with internal fanning bedding dips and alluvial and lacustrine sedimentary facies distributions.
Figure 13. Paleogeographic map for ∼62 Ma showing the new interpretations of the Greymouth and other West Coast basins in the wider context of the West Coast-Taranaki Rift System (updated from Ferguson 1993; Haskell 2007; Barnes and Ghisetti 2013; Dutton et al. 2013; Suggate 2014; Monteith 2015; Strogen et al. 2017).
Extensio nal vs transtensional half-grabens

Asymmetrical half-graben basin geometry filled with alternating fluvial and lacustrine sediments is associated with both transtensional strike-slip and purely extensional basins (Nilsen and Sylvester 1995; Leeder 1999; Gawthorpe and Leeder 2000; Wu et al. 2009), therefore it is possible that the Greymouth Basin could have had an oblique strike-slip component (e.g. Laird 1993, 1994; Bassett et al. 2006). However, the basin geometry for a strike-slip pull-apart basin doesn’t fully match what is observed in the Greymouth Basin; there is no evidence for an eastern sediment source and there is no evidence for shingling of the alluvial fan structures in the superb exposure at TMB in the northwest. In addition, there is no evidence of any large coeval strike-slip faults associated with the West Coast-Taranaki Rift System (Reilly et al. 2015; Strogen et al. 2017) nor from the coeval Late Cretaceous Canterbury Rift Basin (Barrier et al. 2020). It may be that any evidence was overprinted by younger Cenozoic transpression (Browne et al. 2012; Jongens et al. 2012), but no structural evidence has yet been found. In addition, recent geochemical analysis of a basaltic breccia in the Morgan Member indicates tholeiitic composition (Steadman 2018), which is commonly associated with purely extensional rift basins (Tatar et al. 2007; Mathieu et al. 2011). It supports the interpretation of the Greymouth Basin as more probably formed by pure extension. Therefore, although the possibility of a minor strike-slip component cannot be completely ruled out, it was likely to have been insignificant.

Relation to other West Coast sub-basins

There are a number of nearby sub-basins that are included in the West Coast Basin that contain the latest Cretaceous to Paleocene strata; these include the Takutai Half-Graben (already discussed), the Pike River Coalfield to the north of Greymouth Basin, and several smaller, poorly exposed and/or poorly identified basins (Figure 13; Newman 1981, 1985; Carter et al. 1986; Bishop 1992; Ferguson 1993; Haskell 2007; Monteith 2015).

Further south, the Kumara-2 and Arahura-1 wells (Figures 1, 2 and 13) penetrated complete successions of Eocene Brunner Formation and partial successions of Paparoa Formation (Suggate and Waight 1999; Haskell 2007) with minor outcrop further south at Kowhitirangi (Ferguson 1993). Taramakau-1, located between the two wells, encountered Eocene Brunner Formation unconformably overlying the older Pororari Group, likely representing a basement high (Figure 13; Gerling 1964; Haskell 2007). The presence of latest Cretaceous to Paleocene strata on both sides of a basement high suggests the possibility of two local sub-basins south of Greymouth (Figure 13).

The Pike River Coal Field, northeast of Greymouth, contains ~280 m thick package of Paparoa Formation (Ferguson 1993). This basin is crosscut by the Montgomerie-Mount Davy Fault Zone, originally interpreted as the basin bounding fault (Newman 1985; Ferguson 1993). However, the presence of conglomerates in boreholes located in the western part of the coal field, the south-eastward paleoflow direction measured from cross-beds (Newman 1985; Ferguson 1993), and the younger age of the eastern margin Montgomerie-Mount Davy Fault Zone (Suggate 2014) suggest that the basin bounding fault could have been located on the western side of the coal field instead. Paparoa Formation also crops out in the headwaters of the Fox and Pororari rivers and Bullock Creek to the north (Figure 13; Ferguson 1993).

A recent investigation of the stratigraphy in the Buller region further north concludes that some of the sediments that had been previously classified as Eocene Brunner Formation are more likely the equivalent of the Paparoa Formation (Figures 1, 2 and 13; Dutton et al. 2013; Monteith 2015). The Brunner Formation overlies the older sediments with a slight angular unconformity, identified in a difficult to access cliff on Mt Rochfort (Monteith 2015). The section of older conglomerates is ~270 m of clast supported, poorly imbricated, moderately sorted, cobble conglomerate interpreted as deposited by braided rivers on alluvial fans. Imbrication suggests a general SW paleoflow direction. Clasts were primarily derived from local Greenland Group basement. Slightly to the south-east in the Te Kuha sector, coal analyses indicate the presence of low ash, low sulphur coals more similar to the Paparoa coals than to the Brunner (Dutton et al. 2013). Four drill holes penetrated Brunner, Paparoa equivalent, and Hawks Crag Breccia in a complete section representing the two phases of rifting and transgression. About 100 m of fining upward, channelised sandstones have been interpreted as meandering river deposits and assigned to the Paparoa Formation equivalent (Monteith 2015). The ~270 m thick conglomerates at Mt Rochfort and lateral meandering river sandstones at Te Kuha suggest fault-controlled deposition in a half graben, sub-basin of the West Coast Basin (Monteith 2015) providing a geographic link to the better studied Taranaki Basin to the north.

Relation to the wider West Coast-Taranaki Rift System

The Taranaki Basin portion of the West Coast-Taranaki Rift System produced a number of NNE-SSW oriented, en-echelon sub-basins with the greatest accumulations in the Pakawau, Manaia, and Maui...
sub-basins in the southern Taranaki Basin and the Moa and Te Ranga sub-basins to the north (King and Thrasher 1996; Higgs et al. 2010; Reilly et al. 2015). The most southern of the sub-basins is the Pakawau Sub-basin, bound to the east and west by the Wakamarama and Kahurangi faults (Figure 13; Fohrmann et al. 2012; Strogen et al. 2017; Smithies et al. 2020). Both faults were normal faults in the Late Cretaceous which were then reactivated in the Miocene as reverse faults (Reilly et al. 2015). The Pakawau Sub-basin was infilled by the Late Cretaceous Rakopi and North Cape formations (Pakawau Group) and Paleocene Farewell Formation (Kapuni Group) (Bal and Lewis 1994; King and Thrasher 1996; Browne 2009; Higgs et al. 2010). The Rakopi Formation comprises terrestrial coal-bearing, meandering river sandstones interfingering with alluvial fan conglomerates shed off the Wakamarama Fault to the east (Browne et al. 2008; Smithies et al. 2020). A marine transgression, possibly due to increased subsidence, resulted in submergence of the Pakawau Sub-basin to form a tidally dominated, shallow marine embayment with estuarine sub-tidal to salt marsh tidal flat interfingering with gravelly fan delta deposits off the Wakamarama Fault to the east (Browne et al. 2008; Joyce 2018; Smithies 2018). Subsidence in the basin then slowed in the Paleocene resulting in regression and deposition of Farewell Formation meandering river sandstones and coals interfingering with gravelly fans from the east (Smithies 2018). Provenance analysis indicates conglomerates from all three formations were derived from local Takaka Terrane basement rocks uplifted by the Wakamarama Fault (Smithies et al. 2020). There is no evidence of strike-slip faulting from the facies, provenance, or basement outcrop pattern leading to an interpretation of pure extension (Strogen et al. 2017; Smithies et al. 2020). This agrees with earlier work done by Reilly et al. (2015) who found no evidence of strike-slip movement in seismic reflection studies in the offshore southern Taranaki Basin. Our work in the Greymouth Basin also indicates that pure extension with little to no strike-slip component was dominant during the West Coast-Taranaki rift phase further south in New Zealand.

Conclusions

A detailed sedimentary facies analysis of the conglomerates and interbedded fine-grained sediments found on the northwestern side of the Greymouth Rift Basin was used to determine basin geometry and tectonic setting. The Gravely Braided River Facies Association (FA1) is interpreted as deposited by high energy stream flows, subaerial debris flows, and low energy floodplain processes on the lower slopes of streamflow dominated alluvial fans. These interfinger laterally with basin centre, coal-bearing, sandy meandering river systems to the southeast. The Gravely Delta Front Facies Association (FA2) is interpreted as the result of a gravelly delta with mouth bar progradation, channel avulsion, and shallow inter-distributary bays. The Gravely Prodelta Facies Association (FA3) was deposited by high-density turbidity currents and subaqueous debris flows with abundant soft-sediment deformation. These interfinger laterally with the Lacustrine Facies Association (FA4) in the basin centre as low gradient Hjulström-type type fan deltas.

Facies that thicken and coarsen as flows, subaerial debris flows, and low energy streamflows, subaerial debris flows, and low energy floodplain processes on the lower slopes of streamflow dominated alluvial fans. These interfinger laterally with basin centre, coal-bearing, sandy meandering river systems to the southeast. The Gravely Braided River Facies Association (FA1) is interpreted as deposited by high energy stream flows, subaerial debris flows, and low energy floodplain processes on the lower slopes of streamflow dominated alluvial fans. These interfinger laterally with basin centre, coal-bearing, sandy meandering river systems to the southeast. The Gravely Braided River Facies Association (FA1) is interpreted as deposited by high energy stream flows, subaerial debris flows, and low energy floodplain processes on the lower slopes of streamflow dominated alluvial fans. These interfinger laterally with basin centre, coal-bearing, sandy meandering river systems to the southeast. The Gravely Braided River Facies Association (FA1) is interpreted as deposited by high energy stream flows, subaerial debris flows, and low energy floodplain processes on the lower slopes of streamflow dominated alluvial fans. These interfinger laterally with basin centre, coal-bearing, sandy

Acknowledgements

The authors would like to acknowledge the University of Canterbury Mason Trust Fund for financial supports to field trips to Greymouth and the National Core Store in Featherston. The authors are thankful to Brad Field (Senior Scientist at GNS Science, New Zealand) and Dr. Guy Plint (Professor at the Department of Earth Sciences in the Western University, Canada) for their valuable comments on this research.

Disclosure statement

No potential conflict of interest was reported by the author(s).

Funding

This research was funded by the Ministry of Business, Innovation, and Employment of New Zealand (MBIE) through GNS Science-led research program on New Zealand petroleum source rocks, fluids, and plumbing systems (contract CO5X1507), Research Aim 1.4: Discovering our lacustrine petroleum source rocks, fluids, and plumbing systems (contract CO5X1507), Research Aim 1.4: Discovering our lacustrine

Data availability statement

The supplementary data that support the finding of this study and figures are openly available in Zenodo at https://doi.org/10.5281/zenodo.4706562 and in Figshare at https://figshare.com/s/b50cdd99dfe47a15bf9d9.
References

Allen JRL. 1982. Sedimentary structures: their character and physical basis. New York: Elsevier. 1, 663p.

Anadón P, Cabrera L, Julià R, Marzo M. 1991. Sequential arrangement and asymmetrical fill in the Miocene Rubielos de Mora Basin (northeast Spain). In: Anadón P, Cabrera L, Kelts K, editors. Lacustrine facies analysis. International Association of Sedimentologists, Special Publication, 13; p. 257–275.

Bal A, Lewis DW. 1994. A Cretaceous macrotral estuarine-fluvial succession: Puponga Coal Measures in Whanganui Inlet, onshore Pakawau Sub-basin, northwest Nelson, New Zealand. New Zealand Journal of Geology and Geophysics. 37:287–307.

Barnes P, Ghisetti F. 2013. Offshore faulting and earthquake sources, West Coast, South Island. Stage 2. MBIE Envirolink West Coast Regional Council Advice: 1237-WCRC114. NIWA, Wellington, Unpublished Report, 34 p.

Barrier A, Nicol A. 2017. Impact of Late Mesozoic and younger tectonics on half-graben filling and petroleum prospectivity of the Canterbury Basin, New Zealand. AAPG/SEG International Conference and Exhibition, London, England, October 15–18, 2017.

Barrier A, Nicol A, Browne GH, Bassett KN. 2020. Late Cretaceous coeval multi-directional extension in South Zealandia: implications for eastern Gondwana breakup. Marine and Petroleum Geology. 118:104383.

Bassett K, Cody E, Monteith F, Maitra M. 2014. Greyhump coal field: a fault-bounded basin, but which side? In: Holt KA, editor. Abstract Volume, GeoSciences 2014 Conference, 24th – 27th November 2014, Pukekura Raceway and Function Centre, New Plymouth, New Zealand. Geoscience Society of New Zealand Miscellaneous Publication. 139A:6–7.

Bassett K, Ettmuller F, Bernet M. 2006. Provenance analysis of the Paparoa and Brunner Coal Measures using integrated SEM-cathodoluminescence and optical microscopy. New Zealand Journal of Geology and Geophysics. 49(2):241–254.

Bennett MR, Doyle P, Mather AE. 1996. Dropstones: their origin and significance. Palaeogeography, Palaeoclimatology, Palaeoecology. 121(3–4):331–339.

Bhattacharya JP. 2010. Deltas. In James NP, Dalrymple RW (editors). Facies Models 4, Geological Association of Canada. 6:233–264.

Bishop DJ. 1992. Extensive tectonism and magmatism during the middle Cretaceous to Paleocene, North Westland, New Zealand. New Zealand Journal of Geology & Geophysics. 35(1):81–91.

Bishop DJ, Buchanan PG. 1995. Development of structurally inverted basins: a case study from the West Coast, South Island, New Zealand. In: Buchanan JG, Buchanan PG, editors. Basin inversion. Geological Society Special Publication. 88:549–585.

Blair TC. 1999. Sedimentology of the debris-flow-dominated Warm Spring Canyon alluvial fan, Death Valley, California. Sedimentology. 46(5):941–965.

Blair TC. 2002. Alluvial-fan sedimentation from a glacial-outburst flood, Lone Pine, California, and contrasts with meteorological flood deposits. In: Martini IP, Baker VR, Garzon G, editors. Flood and Megaflood Processes and Deposits: Recent and Ancient Examples. International Association of Sedimentologists Special Publication. 32:111–140.

Blair TC, McPherson JG. 2009. Processes and forms of alluvial fans. In: Parsons AJ, Abrahams AD, editors. Geomorphology of desert environments. Dordrecht: Springer; p. 413–467.

Boggs Jr, S. 2006. Continental (terrestrial) environments. In: Boggs S, editor. Principles of sedimentology and stratigraphy. 5th ed. New Jersey: Pearson Education Limited; p. 213–247.

Bowman RG. 1984. Greymouth coalfield report-Part 1 (Geology and coal resources), Ministry of Economic Development, Wellington, New Zealand. New Zealand Coal Resources Survey, Coal Report Series CR 2395, 84 p.

Bowman RG, Caffyn P. 1984. Greyhump Coal Report. Part II, Rapahoe Sector. Ministry of Economic Development, Wellington, New Zealand. New Zealand Coal Resources Survey, Coal Report Series CR 2396, 28 p.

Boyd R. 2002. Spring Creek Mine drilling 2002. Wellington: Ministry of Economic Development. Coal Report Series CR 3102.

Boyd RI, Lewis DW. 1995. Sandstone diagenesis relating to varying burial depth and temperature in Greymouth coalfield, South Island, New Zealand. New Zealand Journal of Geology and Geophysics. 38(3):333–348.

Breeze N, Browne G. 1999. Review of offshore petroleum data, West Coast, South Island, New Zealand. Ministry of Economic Development, Wellington. Unpublished petroleum report PR 2526, 25 p.

Brewer PA, Leeks JL, Lewin J. 1992. Direct measurement of in-channel abrasion processes. In: Walling DR, Bogen J, Day TJ, editors. Erosion and Sediment Transport Monitoring Programmes in River Basins. International Association of Hydrogeological Sciences. 210:21–29.

Browne GH. 2009. First New Zealand record of probable dinosaur footprints form the Late Cretaceous North Cape Formation, northwest Nelson. New Zealand Journal of Geology and Geophysics. 52(4):367–377.

Browne GH, Field BD, Barrett DJA, Jongens R, Bassett KN, Wood RA. 2012. The geological setting of the darfield and christchurch earthquakes. New Zealand Journal of Geology and Geophysics. 55(3):193–197.

Browne GH, Kennedy EM, Constable RM, Raine JI, Crouch EM, Sykes R. 2008. An outcrop-based study of the economically significant Late Cretaceous Rakopi Formation, northwest Nelson, Taranaki Basin, New Zealand. New Zealand Journal of Geology and Geophysics. 51:295–315.

Carter M, Kelly C, Hillyer A, Toomath A, McDowell P. 1986. Kumara-2, a well completion report, Petcorp (Exploration). Ministry of Economic Development, Wellington unpublished petroleum report PR 1183.

Cody EN. 2015. Sedimentology and Hydrocarbon Potential of the Paparoa Coal Measures Lacustrine Mudstones. M.Sc. thesis, University of Canterbury, Christchurch, New Zealand, 217 p.

Collinson JD. 1996. Alluvial sediments. In: Reading H.G., editor. Sedimentary Environments: processes, facies and stratigraphy. 3rd ed. Oxford: Blackwell Science Limited; p. 37–82.

Cook RA, Zhu H, Sutherland R. 1999. Cretaceous–Cenozoic geology and petroleum systems of the Great South Basin, New Zealand. Lower Hutt, New Zealand: Institute of Geological and Nuclear Sciences. Monograph, 20, 188 P.

Duff SW. 1988. Greyhump project feasibility study. Coal resource assessment Rapahoe sector, Greyhump Coal Field. Ministry of Economic Development, Wellington, New Zealand. Greyhump Coal Ltd., Coal Report Series CR 2902, 96 p.

Dutton A, Newman J, Newman J, Pope J, Field A. 2013. The Te Kuha Sector, Buller Coalfield: A revised model. In:
Pleistocene-Holocene Burdur Lake graben, southwestern Anatolia, Turkey. In: Nemec W., Steel R. J, editor. Fan deltas: Sedimentology and tectonic settings. Glasgow: Blackie; p. 186–196.

Kennedy EM. 2003. Late Cretaceous and Paleocene terrestrial climates of New Zealand: leaf fossil evidence from South Island assemblages. New Zealand Journal of Geology and Geophysics. 46:295–306.

Kereszturi Á, Hargitai H, Postma G. 2015. Delta. Encyclopedia of planetary landforms. New York: Springer Science + Business Media; p. 551–560.

King PR, Naish TR, Browne GH, Field BD, Edbrooke SW, Laird MG. 1993. Cretaceous continental rifts: New Zealand segment of Gondwana. In: Cresswell M.M., Vella P, editor. Gondwana five. Proceedings of the 5thInternational Gondwana Symposium, Wellington, New Zealand, 1980. Rotterdam, Netherlands: A.A. Balkema; p. 311–318.

Laird MG. 1993. Cretaceous continental rifts: New Zealand region. In: Ballance P. F, editor. Sedimentary basins of the world. South Pacific sedimentary basins. Amsterdam: Elsevier Science Publishers B.V.; p. 37–49.

Laird MG. 1994. Geological aspects of the opening of the Tasman Sea. In: van der Lingen GJ, Swanson KM, Muir RJ, editors. Evolution of the Tasman Sea Basin: Proceedings of the Tasman Sea Conference, Christchurch, New Zealand, 27–30 November 1992. Rotterdam, Netherlands, A. A. Balkema; p. 1–17.

Laird MG, Bradshaw JD. 2004. The break-up of a long-term relationship: the Cretaceous separation of New Zealand from Gondwana. Gondwana Research. 7(1):273–286.

Larsen PH. 1988. Relay structures in a Lower Permian base-ment-involved extension system, East Greenland. Journal of Structural Geology. 10:3–8.

Leeder MR. 1999. Sedimentology and sedimentary basins: from turbulence to tectonics. In: Leeder M.R, editor. Oxford, UK: Blackwell; 592 p.

Leeder MR, Gawthorpe RL. 1997. Sedimentary models for extensional tilt-block/half-graben basins. In: Coward MP, Dewey JF, Hancock PL, editors. Continental Extensional Tectonics. Geological Society of London Special Publications. 28(1):139–152.

Lowe DR. 1975. Water escape structures in coarse-grained sediments. Sedimentology. 31:749–745.

Maejima W. 1988. Marine transgression over an active alluvial fan: the early Cretaceous Arida formation, Yuasa-Aridagawa Basin, southwestern Japan. In: Nemec W., Steel R. J, editor. Fan deltas: Sedimentology and tectonic setting. Glasgow: Blackie; p. 303–317.

Mahovic O, Dyvik H, Hammer E. 2018. Diagenetic influence on reservoir quality evolution, examples from Triassic conglomerates/arenites in the Edvard Grieg field, Norwegian North Sea. Marine and Petroleum Geology. 93:247–271.

Maitra MK. 2020. Facies analysis, source rock geochemistry, tectonic evolution, and sequence stratigraphy of the Greymouth Basin, South Island, New Zealand. PhD thesis. University of Canterbury, Christchurch, New Zealand, 313 p.

Maizels J. 1993. Lithofacies variations within sandur deposits: the role of runoff regime, flow dynamics and sediment supply characteristics. Sedimentary Geology. 85:299–325.

Makaske B, Smith DG, Berendsen HJ. 2002. Avulsions, channel evolution and floodplain sedimentation rates of the anastomosing upper Columbia River, British Columbia, Canada. Sedimentology. 49(5):1049–1071.

Maltman AJ, Bolton A. 2003. How sediments become mobilized. In: Van Rensbergen P, Hillis RR, Maltman AJ, Morley CK, editors. Subsurface Sediment Mobilization. Geological Society of London Special Publications. 216:9–20.

Marnette EC, Van Breenen N, Hordijk KA, Cappenberg TE. 1993. Pyrite formation in two freshwater systems in the Netherlands. Geochimica et Cosmochimica Acta. 57(17):4165–4177.

Martinsen OJ. 1989. Styles of soft-sediment deformation on a Namurian (Carboniferous) delta slope, western Irish Namurian Basin, Ireland. In: Whatley MKG, Pickering KT, editors. Deltas: Sites and Traps for Fossil Fuels. Geological Society of London Special Publication. 210:167–177.

Martins-Neto M, Catuneau O. 2010. Rift sequence stratigraphy. Marine and Petroleum Geology. 27(1):247–253.

Mathieu L, de Vries BVW, Pilato M, Troll VR. 2011. The interaction between volcanoes and strike-slip, transtensional and transpressional fault zones: analogue models and natural examples. Journal of Structural Geology. 33(5):898–906.

McPherson JG, Shanmugam G, Moiola RJ. 1988. Fan deltas and braid deltas: conceptual problems. In: Nemec W., Steel R. J, editor. Fan deltas: sedimentology and tectonic settings. London: Blackie; p. 14–22.

Metzger P, Largeau C. 2005. Botryococcus braunii: a rich source for hydrocarbons and related ether lipids. Applied Microbiology and Biotechnology. 66(5):486–496.

Miall AD. 1996. The geology of alluvial deposits. New York: Springer-Verlag. 582p.

Miall AD. 2010. Alluvial deposits. In: James NP, Dalrymple RW, editors. Facies Models 4. St. John’s (Newfoundland): Geological Association of Canada; p. 105–137.

Mills PC. 1983. Genesis and diagnostic value of soft-sediment deformation structures—a review. Sedimentary Geology. 35(2):83–104.

Mohnhoff D, Crouch EM, Naeher S, Sykes R. 2017. Understanding petroleum source rock properties of mid-Cretaceous to Paleocene lacustrine mudstones in New Zealand basins. New Zealand Petroleum Conference, Wellington, New Zealand.

Mohrig D, Heller PL, Paola C, Lyons WJ. 2000. Interpreting avulsion process from ancient alluvial sequences: Guadalope-Matarranya system (northern Spain) and Wasatch Formation (western Colorado). Geological Society of America Bulletin. 112(12):1787–1803.

Monteith FD. 2015. Late Palaeocene-Eocene tectonic-sedimentary evolution of North Westland, Soult Island: An analysis of the Brunner Coal Measures and their basal contact. M.Sc. thesis, University of Canterbury, Christchurch, New Zealand, 147 p.

Mortimer N. 2004. New Zealand’s geological foundations. Gondwana Research. 7(1):261–272.

Mortimer N, Campbell HJ, Tulloch AJ, King PR, Stagpoole VM, Wood RA, Rattenbury MS, Sutherland R, Adams CJ, Collot J, Seton M, Wood RA, Rattenbury MS, Bland KJ, Barrett DJA, Bache F, Edbrooke SW. 2014. High-level
stratigraphic scheme for New Zealand rocks. New Zealand Journal of Geology and Geophysics. 57(4): 402–419.

Mortimer N, Tulloch AJ, Spark RN, Walker NW, Ladley E, Allibone AH, Kimbrough DL. 1999. Overview of the Median Batholith, New Zealand: a new interpretation of the geology of the Median Tectonic Zone and adjacent rocks. In: Storey BC, Rubbridge BS, Cole DJ, de Wit MJ, editors. Gondwana 10: Event stratigraphy of Gondwana: proceedings volume 1. Oxford (England): Elsevier. Journal of African Earth Sciences 29: 257–268.

Nathan S. 1978. Sheet 44 — gyremouth. Geological map of New Zealand 1:63 360. Wellington, New Zealand: Department of Scientific and Industrial Research.

Nathan S, Anderson HJ, Cook RA, Herzer R, Hoskins R, Rattenbury MS, Suggate RP, compilers. 2002. Nemec W, Steel RJ.1984. Alluvial and coastal conglomerates: remarks on terminology and classification. In Colella A, Prior DB, editors. Coarse-grained deltas, International Association of Sedimentologists Special Publication. 10:3–12.

Nemec W, Steel RJ. 1984. Alluvial and coastal conglomerates: their significant features and comments of gravelly mass-flow deposits. In Koster EH, Steel RJ, editors. Sedimentology of Gravels and Conglomerates. Canadian Society of Petroleum Geologists, Memoir. 10:1–31.

Newman J. 1981. Development of the Late Cretaceous–Paleocene Paparoa Coal Measure Basin at Greymouth, the interrelationship of differential subsidence, sedimentary facies, coal seam distribution and coal seam character. Report to Lime and Marble Ltd. for Mines Division, Ministry of Energy, New Zealand.

Newman J. 1985. Palaeoenvironments, coals properties, and their inter-relationships in Paparoa and selected Brunner Coal Measures on the West Coast of the South Island. PhD thesis, University of Canterbury, Christchurch, New Zealand, 269p.

Newman J. 1987. Coal Geology Report Series Volume 3—Coal types and palaeoenvironments in the Rapahoe Sector, Greymouth Coal Field. Ministry of Economic Development, Wellington, New Zealand. Ministry of Energy, Coal Report Series CR 3111, 28 p.

Newman J, Newman NA. 1992. Tectonic and palaeoenvironmental controls on the distribution and properties of Upper Cretaceous coals on the West Coast of the South Island, New Zealand. In: McCabe PJ, Parish JT, editors. Controls on the distribution and quality of cretaceous coals. Boulder, Geological Society of America Special Paper. 267:347–368.

Nilsen TH, Sylvester AG. 1995. Strike-slip basins. In: Busby C.J., Ingersoll R.V., editor. Tectonics of sedimentary basins. USA: Blackwell Science, Cambridge; p. 425–457.

Nunweek CN. 2001. Depositional controls on peat accumulation and coal characteristics, Dunollie and Brunner Coal Measures, Southern Rarrahoe Sector, Greymouth. M.Sc. thesis, University of Canterbury, Christchurch, New Zealand, 161 p.

Owen G. 1996. Experimental soft-sediment deformation: structures formed by the liquefaction of unconsolidated sands and some ancient examples. Sedimentology. 43 (2):279–229.

Owen G, Moretti M. 2011. Identifying triggers for liquefaction-induced soft-sediment deformation in sands. Sedimentary Geology. 235(3–4):141–147.

Peacock DCP, Sanderson DJ. 1991. Displacements, segment linkage and relay ramps in normal fault zones. Journal of Structural Geology. 13:721–733.

Postma G. 1984. Mass-flow conglomerates in a submarine canyon: Abrioja fan-delta, Pliocene, southeast Spain. Sedimentology of Gravels and Conglomerates. 10:237–258.

Postma G. 1990. Depositional architecture and facies of river and fan deltas: a synthesis. In: Colella A, Prior, DB, editors. Coarse-Grained Deltas. International Association of Sedimentologists Special Publication. 10:13–28.

Postma G. 2003. Fan delta. In: Middleton GV, Chruch MJ, Coniglio M, Hardie LA, Longstaffe FJ, editors. Encyclopedia of sediments and sedimentary rocks. Boston: Kluwer Academic Publishers; p. 272–274.

Postma G, Babic L, Zupanic J, Roe SL. 1988. Delta-front failure and associated bottomset deformation in a marine, gravelly Gilbert-type fan delta. In: Nemec W., Steel RJ., editor. Fan deltas. Sedimentology and tectonic settings. Glasgow and London: Blackie; p. 91–102.

Prior DB, Yang ZS, Bornhold BD, Keller GH, Lu NZ, Wiseman WJ, Wright LD, Zhang J. 1986. Active slope failure, sediment collapse, and sill flows on the modern subaqueous Huanghe (Yellow River) delta. Geo-Marine Letters. 6(2):85–95.

Raine I, Kennedy L, Ciowes C. 2017. A vegetation and paleoclimate record from New Zealand Cretaceous coal measures. NZ Petroleum Geoscience Workshop, Wellington, New Zealand, September 28–29.

Raine JI. 1984. Outline of a Palynological Zonation of Cretaceous to Paleogene Terrestrial Sediments in West Coast Region, South Island, New Zealand. New Zealand Geological Survey Report 109. 82 p.

Rattenbury MS, Isaac MJ. 2012. The QMAP 1:250 000 Geological Map of New Zealand project. New Zealand Journal of Geology and Geophysics. 55(4):393–405.

Reading HG. 1996. Sedimentary environments: processes, facies and stratigraphy. Third Edition. Oxford: Wiley-Blackwell. 704 p.

Rees C, Palmer J, Palmer A. 2018. Gilbert-style Pleistocene fan delta reveals tectonic development of North Island axial ranges, New Zealand. New Zealand Journal of Geology and Geophysics. 61(1):64–78.

Reilly C, Nicol A, Walsh JJ, Seebeck H. 2015. Evolution of faulting and plate boundary deformation in the Southern Taranaki Basin, New Zealand. Tectonophysics. 651–652:1–18.

Renaut RW, Gierlowski-Kordesch EH. 2010. Lakes. In: James NP, Dalrymple RW, editors. Facies Models 4. St. John’s (Newfoundland): Geological Association of Canada; p. 541–575.

Roberts S, Jackson JA. 1991. Active normal faulting in central Greece: an overview. In: Roberts AM, Yielding G, Freeman B, editors. The Geometry of Normal Faults. Geological Society London Special Publication. 56:125–142.

Rosendahl BR. 1987. Architecture of continental rifts with special reference to East Africa. Annual Review of Earth and Planetary Sciences. 15(1):445–503.

Schoonen MA. 2004. Mechanisms of sedimentary pyrite formation. Geological Society of America Special Papers. 379:117–134.
