Field Trip 2: Landuse and stratigraphy within the eastern Whanganui Basin

North Island, New Zealand

View east across the Rangitikei Valley, taken from a small terrace remnant off Iron Works Road, Mangaonoho.

Callum Rees[1], Alan Palmer[1], Julie Palmer[1], and Malcolm Todd[2]

[1]*Environmental Sciences, School of Agriculture and Environment, Massey University, Palmerston North, New Zealand (C.Rees@massey.ac.nz)*

[2]*Horizons Regional Council, Palmerston North, New Zealand*

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Welcome

Welcome to the Manawatū! It is fantastic to be a part of the Geoscience Society of New Zealand Annual Conference 2022 after being setback by COVID-19 related difficulties last year. We will be setting off on an adventure to look at some of the nearby yet unfrequented sections of the Manawatū- Rangitikei area. The first two stops will be on lesser-travelled gravel roadsides. Caution must be taken as stock and forestry trucks sometimes use these roads. Field trip participants will be issued with high visibility jackets. You will be able to hear oncoming traffic so please listen to instructions when asked to move off to the side of the road. The sediments we are going to look at are loose and cliffs are prone to soil slip, please keep this in mind and be cautious while examining sections. The field trip leader will have a personal locator beacon and will be identified prior to beginning the field trip. First aid kits will be in the vehicles. If you have any allergies, medical conditions or special considerations please alert one of the field trip leaders at the start of the day or earlier. Any issues will be dealt with discretely and we will have emergency backup provision to transport you back to Palmerston North if required.

Setting the scene

The Manawatū-Whanganui Region contains geologically young and dynamic landscapes that have largely emerged from the sea over the last 1 Ma (Beu et al. 1981; Shane 1991). The variety of relatively young soils and landforms provide valuable natural resources and also pose environmental challenges for local communities and society. Over the last 150+ years, widespread clearance of the land and establishment of pastoral farming has led to accelerated erosion and increased sediment and nutrient loads within the regions freshwater environments. The impact of landuse change is exacerbated by the regions rock types, which are susceptible to episodes of widespread shallow and deep-seated landsliding, causing disruption to primary production, infrastructure and development projects (Crozier & Pillans 1991; Hancox & Wright 2005; Massey 2010; Williams et al. 2021).

The regions active landscapes are highlighted by permanent closure and replacement of the Manawatū Gorge Road with the $620 million Te Ahu a Turanga - Manawatū Tararua Highway, due to ongoing instability and landslides. Our first stop will take us north of Palmerston North to the lower Pohangina Valley and Manawatū saddle, where the proposed new road begins its journey east toward Woodville. Here we will pause to explore some Mid Quaternary volcaniclastic sediments in a farm road cutting and get a feel for the age and nature of the sedimentary deposits preserved on the western flank of the Ruahine Range. This area contains evidence of a shallow marine seaway called the Manawatu strait that once connected the Whanganui Basin with the east coast basin across what is now the main axial range of New Zealand between c. 5-1.6 Ma (Rees et al. 2018c).

Since the 1940s gully erosion within the weakly consolidated to unconsolidated lithologies of the Pohangina Valley have prompted investigation into land use management and erosion control, including establishment of the Te Awa experimental farm on Coulters Line (Miri 1999). This work promotes vegetation for erosion control, including gully retirement, space planting and afforestation. Our trip will take us west from the foot of the Manawatū saddle to the western limb of the Pohangina Anticline at Finnis Road, where we may observe the geologically young succession (c. 1-1.2 Ma) and discuss how this relates to some of the worst gully erosion within the Horizons Region.

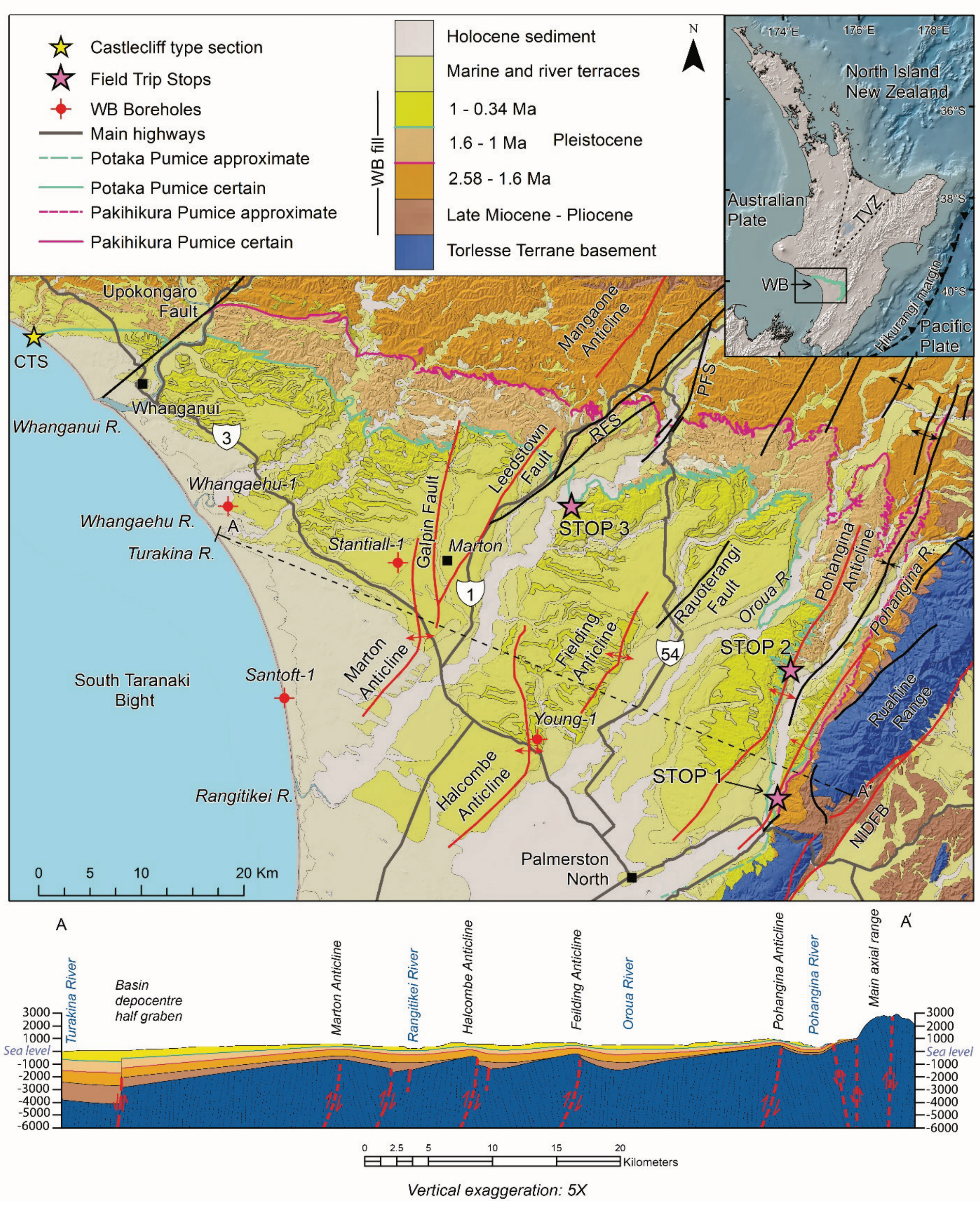

Fig. 1: Map displaying the onland portion of the Whanganui Basin (WB) relative to the Taupo Volcanic Zone (TVZ) and Hikurangi margin. Field trip sites and section locations are shown, including the Castlecliff type section (CTS), North Island Dextral Fault Belt (NIDFB), Rangitikei Fault System (RFS) and the Pakihikura Fault System (PFS). A simplified cross section A to A' illustrates the undulating basement surface beneath the Late Miocene - Pleistocene basin fill, constrained by available borehole data. Data sourced from Feldmeyer et al. (1943); Te Punga (1952); Fleming

(1953): Milne (1968); Seward (1976); Anderton (1981); Melhuish et al. (1996), Pillans et al. (2005); GNS Science and Rees et al. (2019).

The geology on this field trip is dominated by the marginal marine to shallow marine sedimentary fill of the Whanganui Basin (WB) (Fig. 1), which provides one of the most complete records of Quaternary climate change, exposed onland, anywhere in the world (Naish et al. 1998). Nowhere is this basin fill more spectacularly exposed than in the Rangitikei River valley, renowned for its steep river cliffs and beautiful terraced landscapes.

Our last site, Stop 3, will take us up a small side tributary of the Rangitikei River where evidence of past volcanism, sea level change and uplift may be observed. Our stroll up the stream valley will take us through an array of vertically stacked, fossiliferous sedimentary deposits. These units reflect a range of laterally adjacent depositional environments across the paleo-shelf shoreline transect at around c. 0.9 Ma.

Whanganui Basin

Whanganui Basin (WB) in the lower North Island of New Zealand (Fig. 1 and 2) contains c. 4500 m of marginal to shallow marine sedimentary deposits, providing a rich archive of Quaternary environmental change (Anderton 1981; Pillans 1994). The basin is located 200 km west of the Hikurangi Trough and is thought to have formed through a combination of lithospheric loading and compressional downwarping driven by coupling of the Pacific and Australian plates (Davey 1977; Stern et al. 1993; Pillans 2017). The basin depocentre migrated south through the Plio-Pleistocene together with simultaneous uplift along the northern and eastern basin margins. Ongoing uplift and fluvial incision have resulted in excellent on-land exposures through the gently southward dipping basin succession (Anderton 1981; Pillans 2017). The basin fill provides a record of eustatic sea-level fluctuations, volcanic events, and phases of regional subsidence and uplift related to the paleogeographic development of the lower North Island (Pillans 1994; Proust et al. 2005).

Fluvio-estuarine facies within WB coastal plain settings are conducive to the preservation of organic matter and volcanic products, evidenced by abundant lignite, leaf, wood, fossil and frequent tephra layers (Te Punga 1952; Milne 1968; Brackley 1999; Rees et al. 2018b). Spatio-temporal changes in both depositional style and extent of preservation become evident within the WB succession between the coastal type section near Whanganui and the present day axial range (Fleming 1953). A particularly long-lived 500 ka unconformity at the base of the Butlers Shell Conglomerate (base of Okehu Group), along the Whanganui coast, is not present to the same extent within the eastern WB succession, replaced by a more complete volcaniclastic record (Abbott 1994; Pillans et al. 2005). The eastern WB succession has only been lightly explored and presents difficulties in correlation to the central and western basin succession due to deposition at a basin margin characterised by frequent subaerial exposure and erosion (Naish et al. 2005).

The Pleistocene succession of the WB is characterised by siliciclastic to bioclastic sediments deposited within coastal plain, shoreface and shelf environments during climatically controlled sea-level cycles (Abbott 1998; Carter & Naish 1998; Abbott & Carter 1999). The cyclothemic strata exposed along the Whanganui coastline and documented by Fleming (1953) and Abbott (1994) among others (Saul et al. 1999; Naish et al. 2005) were deposited within the late rise, high stand and early fall of sea level, with low stand periods characterised by erosional surfaces formed during subaerial exposure of the continental shelf (Naish et al. 1998). Some WB sequences, particularly within the eastern basin, contain regressive and low stand subaerial surfaces including soils, lignite layers and coastal plain facies (Abbott 1998; Rees et al. 2019). Marginal marine to coastal plain facies of the WB present relatively unexplored archives of volcanic and environmental signals (Mildenhall 1975; Russell 1984, 1986; Russell & Mildenhall 1990; Russell 1993; Naish et al. 2005; Pillans et al.

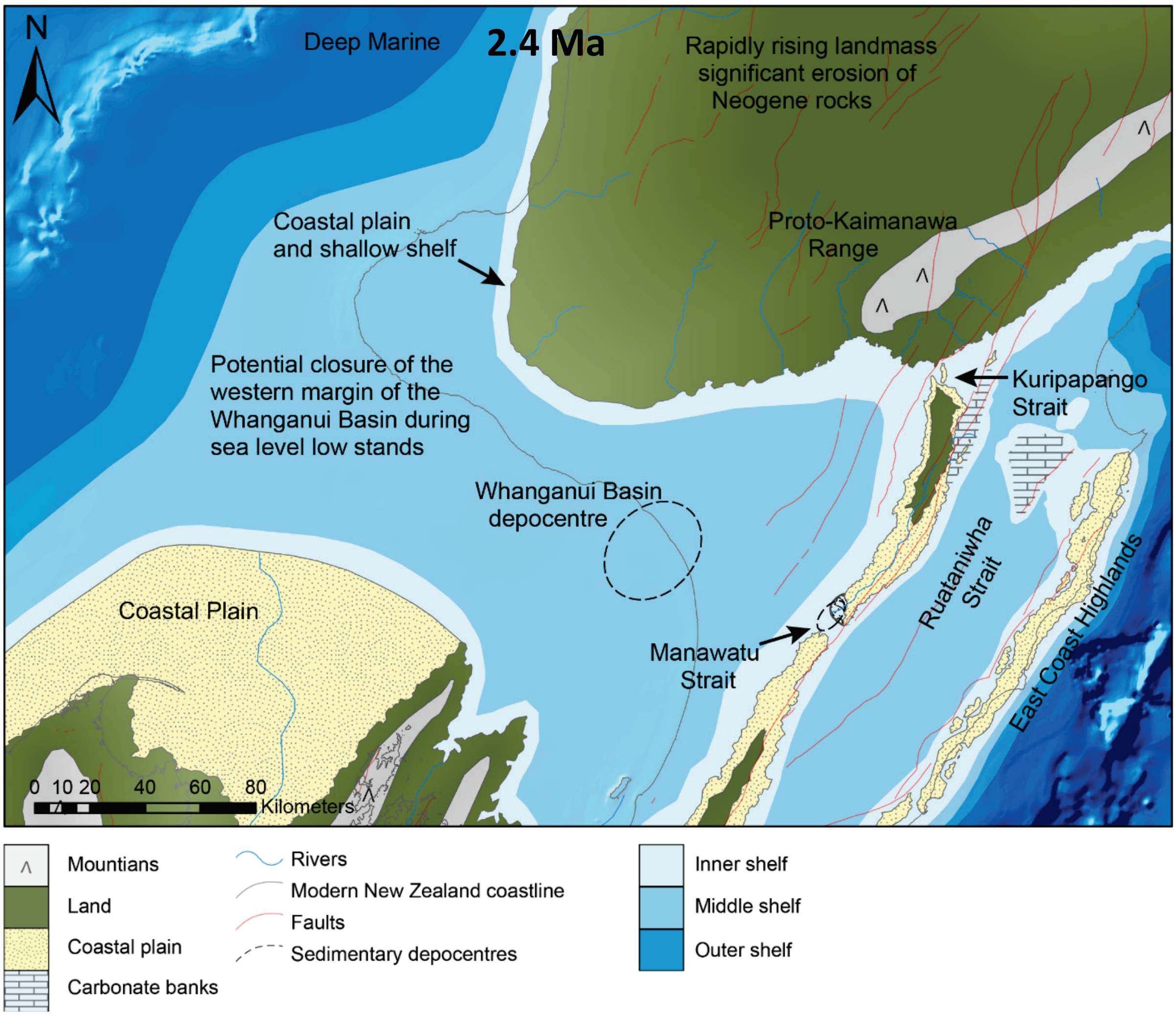

Fig. 2: Schematic of the lower North Island during a sea level high stand in Early Nukumaruan time (2.4 Ma). Sources of information include Beu (1995); King and Thrasher (1996); Field et al. (1997); Browne (2004); Kamp et al. (2004); Pulford and Stern (2004); Bland (2006); Kamp and Furlong (2006); Townsend et al. (2008); Bunce et al. (2009); Nicol (2011); Trewick and Bland (2012).

Taupo Volcanic Zone

Rhyolitic volcanism is thought to have begun in the Taupo Volcanic Zone (TVZ) at around 1.6 Ma (Houghton et al. 1995; Pillans et al. 2005; Wilson et al. 2009). Since c. 1.6 Ma at least 34 ignimbrite sheets have been erupted during explosive volcanic events, resulting in large influxes of volcaniclastic sediments into peripheral sedimentary basins (Wilson et al. 1984; Alloway et al. 2005). Initial volcanic activity originated from the Mangakino caldera (Fig. 3), during two periods of caldera forming activity from 1.68-1.53 and 1.21-0.95 Ma (Briggs et al. 1993; Houghton et al. 1995; Krippner et al. 1998). Voluminous widespread welded and non-welded ignimbrites and tephra-fall deposits originating from the Mangakino Volcanic Centre provided an important source of sediment to the WB during Castlecliffian time (Shane 1993; Shane et al. 1996)

Each influx of geochemically discrete tephra-fall and/or volcaniclastic sediment into the succession has been used to aid in development of the WB chronostratigraphic framework, providing extremely useful horizons for field mapping (Feldmeyer et al. 1943; Te Punga 1952; Seward 1974; Pillans et al. 2005). Tephrostratigraphy augments biostratigraphic and magnetostratigraphic records to provide accurate correlation and dating of the basin succession (Pillans et al. 1994; Turner et al. 2005; Beu 2012).

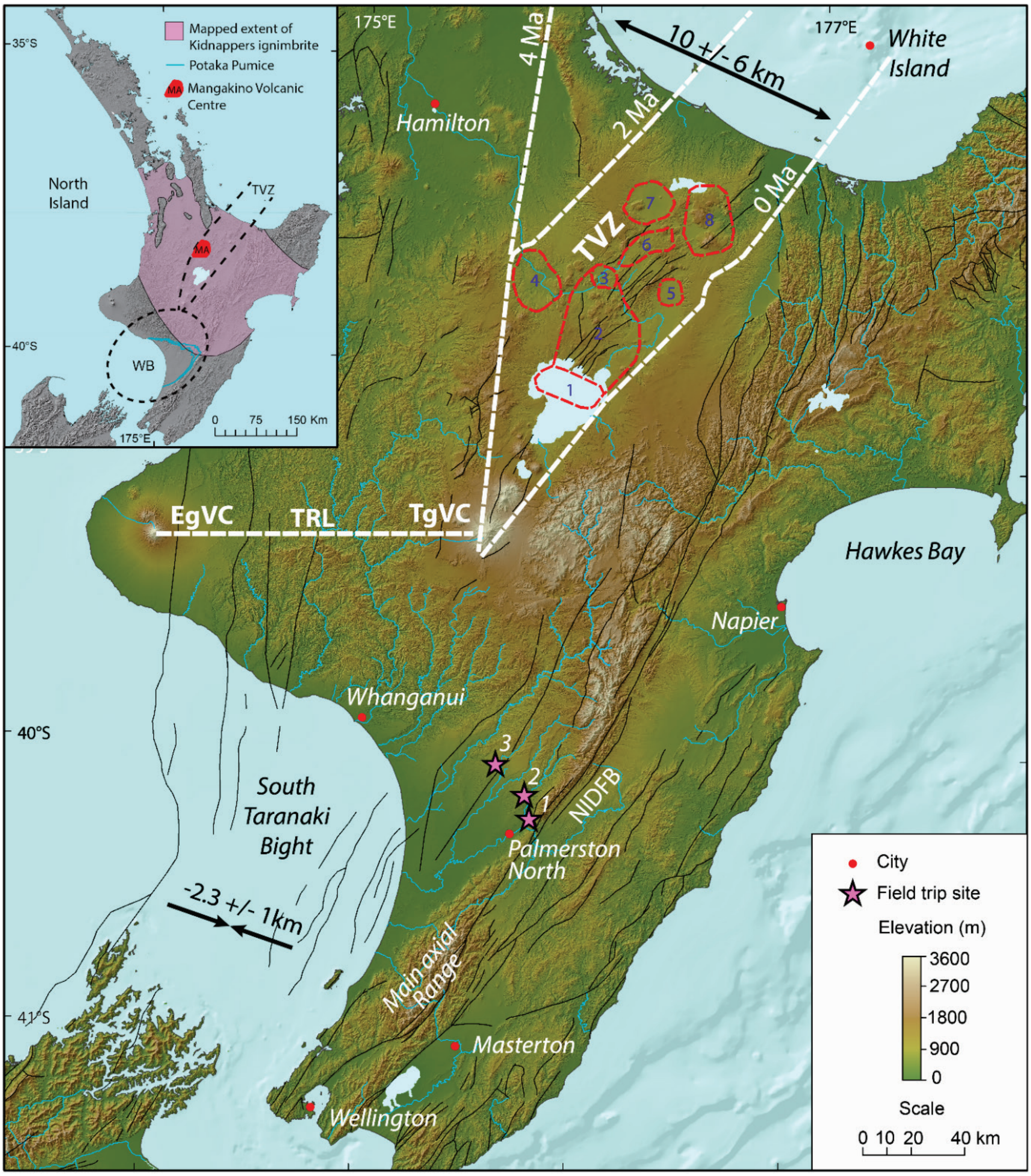

Fig 3: Map of the North Island showing the field trip sites in relation to the Whanganui Basin (WB) and the Taupo Volcanic Zone (TVZ). Red outlines represent the eight major calderas within the TVZ: 1 = Taupo, 2 = Whakamaru, 3 = Ohakuri, 4 = Mangakino, 5 = Reporoa, 6 = Kapenga, 7 = Rotorua, 8 = Okataina. TgVC = Tongariro Volcanic Centre, EgVC = Egmont Volcanic Centre, TRL = Taranaki Ruapehu Line. Black arrows indicate Pliocene shortening and extension (Nicol et al. 2007). Black lines indicate faults, including the North Island Dextral Fault Belt (NIDFB). Our three field sites are indicated by pink stars.

Pakihikura Pumice (c. 1.6 Ma)

Geologists of the Superior Oil Company (Feldmeyer et al. 1943) originally mapped the Pakihikura Pumice as the 'Basal Castlecliffian Ash' from the lower Pohangina Valley, across the basin to Makirikiri Valley 9.5 km NE of Whanganui City, a feat and level of detail unmatched by any study since. Te Punga (1952) originally introduced the name Pakihikura Pumice during his comprehensive mapping of the Rangitikei Valley, a study that early on was conducted for the Superior Oil Company before later submission as a PhD thesis at Victoria University of Wellington and publication in the New Zealand Geological Survey Memoir 8, 1952 (Beu, A.G. Pers. coms. 2018). Fleming (1953) mapped an equivalent pumiceous horizon, the Makirikiri Tuff Formation, from Makirikiri Valley to Turakina Valley (Fig. 4), correlating it with the Basal Castlecliffian Ash of Feldmeyer et al. (1943). Seward (1976) summarised past nomenclature, recognising at least three distinct, separate tephra-fall and/or volcaniclastic deposits within the Makirikiri Tuff Formation, based on ferromagnesian mineral assemblages. The three deposits where named as Pakihikura Pumice, Ridge Tephra and Mangapipi Tephra (Table 1). This work formalised the Pakihikura Pumice as a member within the Makirikiri Tuff Formation of Fleming (1953).

Fig. 4: Pakihikura Pumice exposed at a cutting on Turakina Valley Road (BL33 0135 7932).

Te Punga (1952) described a 25 m section of tephric beds at Pakihikura Bluff on the eastern side of the Rangitikei River, naming it the Pakihikura Pumice. Boellstorff and Te Punga (1977) defined the type locality of the Pakihikura Pumice as a road cutting 200 m south of Pakihikura Stream. Seward (1976) revisited the type section, describing the unit as a white vitric ash 0.5 m thick interbedded within tephric sandstone. In the Turakina and Rangitikei River valleys the Pakihikura Pumice occurs within a 100 to 150 m thick sequence of marginal marine to non-marine carbonaceous mudstone and tephric sandstone (WB Cycle 22) (Pillans et al. 2005). In the lower Pohangina Valley, the Pakihikura Pumice occurs as a 0.5 m thick, white, tephric bed in a succession of alternating tephric sandstone and carbonaceous mudstone including common lignite, within Waitokanui Member, Takapari Formation (Carter 1972; Rees et al. 2018b).

The source of the Pakihikura Pumice is uncertain, however, two ignimbrites named Ngaroma Ignimbrite and ignimbrite Unit-B erupted from the Mangakino caldera have $^{40}Ar/^{39}Ar$ ages consistent with the zircon and glass fission track ages of the Pakihikura Pumice (Wilson et al. 1986; Briggs et al. 1993; Houghton et al. 1995; Wilson et al. 2009). Ages of 1.55 ± 0.05 Ma for Ngaroma Ignimbrite and 1.53 ± 0.04 Ma for ignimbrite Unit-B are presented by Wilson et al. (2009). Difficulties arise from limited exposure, partial welding and vapour phase alteration of the ignimbrite sheets, resulting in an inconclusive match with their suspected volcaniclastic equivalent. Alloway et al. (1993) presented an isothermal plateau fission track (ITPFT) age of 1.63 ± 0.15 Ma for the Pakihikura Pumice during revision of the WB chronostratigraphic framework. Pillans et al. (2005) presented an ITPFT age of 1.58 ± 0.08 Ma for the Pakihikura Pumice, correlating it to the upper regressive part of Cycle 22, late MIS 55 and early MIS 54.

Pakihikura Pumice is a particularly important bed for mapping in the WB, largely due to its striking character in outcrop and unique geochemistry (Fig. 4 and 5). It signals a prominent change in the WB sedimentary succession, after which point deposits become highly tephric and carbonaceous with shellbeds dominated by *Austrovenus stutchburyi* and *Barytellina crassidens*. This change represents initiation of rhyolitic volcanic activity within the TVZ (Alloway et al. 1993; Krippner et al. 1998; Briggs et al. 2005; Allan et al. 2008) and its influence on the WB.

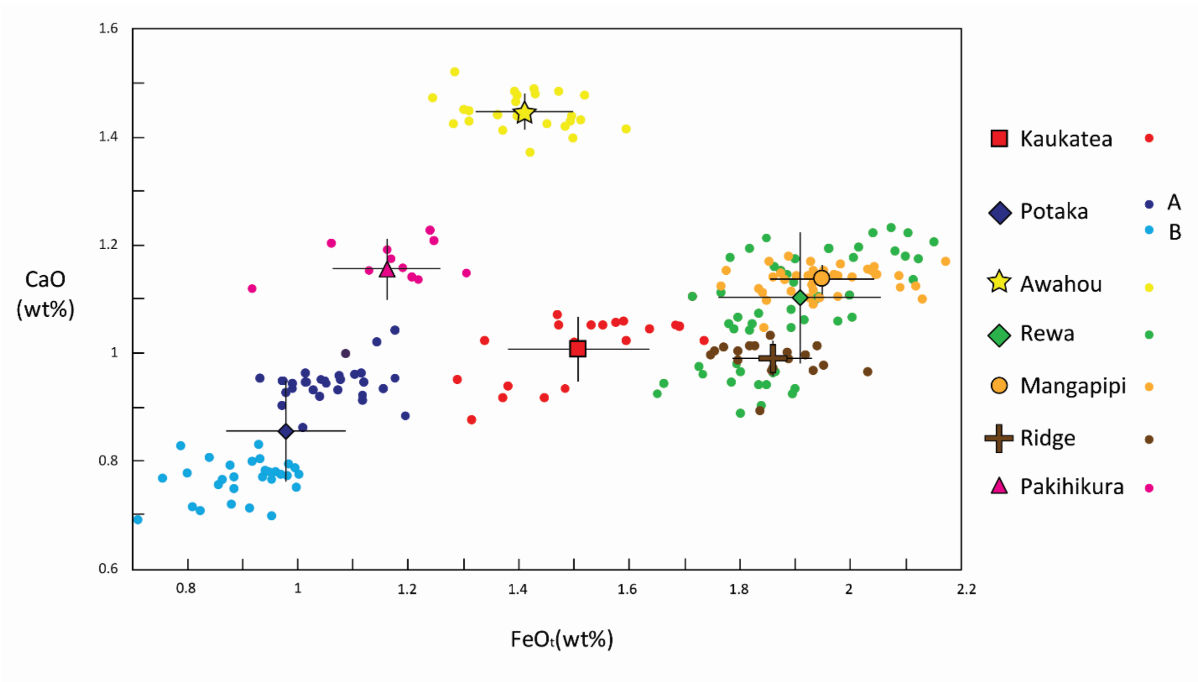

Fig. 5: CaO versus FeO_t (wt%) composition of glass shards analysed from seven tephra-fall and/or volcaniclastic deposits in the lower Pohangina Valley. Mean analyses are displayed together with error bars representing one standard deviation. Note the Potaka dataset is separated into two populations A and B. Population A represents the initial eruptive phase that was enriched in FeO_t and CaO.

Potaka Pumice (c. 1 Ma)

Feldmeyer et al. (1943) mapped the Potaka Pumice as the Kimbolton Ash Horizon tracing its extent from Pohangina Valley east to the Rangitikei Valley (Fig. 6 and 7). Potaka Pumice was originally named by Te Punga (1952) during mapping of the Rangitikei Valley, describing a strongly current-bedded pumiceous unit with abundant rounded pumice pebbles. Seward (1976) summarised existing nomenclature, correlating the Potaka Pumice to the Kaimatira Pumice Sand of Fleming (1953). A critical change has occurred since this time, due to extensive work on the volcaniclastic succession of the WB, summarised in Pillans et al. (2005). Namely, the Kaimatira Pumice Sand is now defined as the first influx of Potaka Pumice into the WB (Pillans et al. 2005), resulting in exclusion of the Rewa Pumice and as a result an emendation of the definition by Seward (1976).

The Potaka Pumice is a widespread marker horizon used to correlate between Pleistocene marine and non-marine facies for a distance of around 600 km (Alloway et al. 1993; Shane 1994). Its occurrence within the Jaramillo Subchron, close to the paloemagnetic transition from normal to reversed polarity, makes it a valuable chronostratigraphic marker bed. Alloway et al. (1993) presents an ITPFT age of 1.05 ± 0.05 Ma for the Potaka Pumice. Naish et al. (1998) correlate Potaka Pumice to the base of Cycle 34 (MIS 27), providing an astronomical age of 0.99 Ma.

Fig. 6: Ripples in Potaka Pumice exposed in a side tributary of the Rangitikei River (BL34 1575 6854).

The Potaka Pumice has been correlated to two separate ignimbrites; the Kidnappers and Rocky Hill ignimbrites, erupted from the Mangakino Caldera (Wilson et al. 1995; Alloway et al. 2005; Wilson et al. 2009; Cooper et al. 2012). The Kidnappers eruption generated a fine-grained phreatomagmatic fall deposit followed by an exceptionally widespread, non-welded ignimbrite (Wilson et al. 1995; Cooper et al. 2017). After a brief interval of erosion, the Rocky Hill eruption occurred, generating a partly-welded ignimbrite (Cooper et al. 2016). The two deposits yield identical $^{40}Ar/^{39}Ar$ ages of c. 1.0

Ma (Wilson et al. 1995). Following ignimbrite emplacement, paleo-fluvial systems draining the central North Island eroded and transported Potaka pumice into peripheral sedimentary basins on both sides of the present day main axial range, providing evidence on the timing of uplift and subsequent formation of the ranges (Shane 1991; Pillans 1994; Shane et al. 1996).

The first influx of Potaka Pumice into the WB sedimentary record defines the base of the Kai Iwi Group at the Kaimatira Pumice Sand Formation. Kaimatira Pumice Sand Formation is able to be traced from the coastal type section to the Rangitikei Valley (Te Punga 1952; Fleming 1953; Seward 1976; Abbott 1994) and Pohangina Valley (Feldmeyer et al. 1943; Milne 1968; Brackley 1999; Pillans et al. 2005; Rees et al. 2018b). The first occurrence of Potaka Pumice has been used to define the upper boundary of the Takapari Formation of Carter (1972) and lower boundary of the Kaimatira Pumice Sand Formation in the lower Pohangina Valley (Rees et al. 2018c). It also defines the boundary between the Mangatarata and Mangahao formations in the Dannevirke and Woodville areas on the eastern side of the main axial ranges (Lillie 1953; Piyasin 1966). A key characteristic of Potaka Pumice is the occurrence of compositional variation within its glass-shard based major element geochemistry, reflecting variation in the eruptive phase (Shane 1994). A segregation of compositions indicates that glass of the initial eruptive phase was enriched in FeO and CaO (see populations A and B, Fig. 5).

Fig 7: Potaka Pumice in the lower Pohangina Valley (BM35 3611 3800).

Group	Formation	Member	Tephra-fall and/or volcaniclastic deposit	ITPFT age (Ma)	Zircon FT age (Ma)	Astronomical age (Ma)	U-Th-Pb and U-Pb age (Ma)
Shakespeare	Mangatapu	Rangitawa Pumice	Rangitawa	0.34 ± 0.03		0.35	
	Tokorangi		Onepuhi			0.57	
Kai Iwi	Reu Reu						
	Waiomio	Waiomio Shellbed	Kupe	0.63 ± 0.08		0.65	
		Waitapu Shell Conglomerate	Kaukatea	0.86 ± 0.08		0.90	
	Kaimatira Pumice Sand	Potaka Pumice	Potaka	1.00 ± 0.03	0.97 ± 0.04	0.99	
Okehu	Rewa	Rewa Pumice	Rewa	1.20 ± 0.14	1.00 ± 0.06	1.19	
	Makirikiri Tuff	Mangapipi	Mangapipi	1.51 ± 0.16	1.26 ± 0.08	1.54	
			Ridge			1.56	
		Pakihikura Pumice	Pakihikura	1.58 ± 0.08	1.66 ± 0.07	1.60	
Maxwell	Upper Maxwell		Birdgrove			1.60	
			Mangahaou			1.63	
			Maranoa			1.64	
			Ototoka	1.72 ± 0.32		1.64	
			Table Flat	1.71 ± 0.12		1.65	
	Rataiti						
Rangitikei	Vinegar Hill Formation		Vinegar Hill	1.75 ± 0.20	1.73 ± 0.08	1.75	
		Waipuru Shellbed	Waipuru	1.79 ± 0.15		1.83	
	Mangaonoho	Mangamako Shellbed	Mangamako			2.05	
	Orangipongo	Ohingaiti Sand	Ohingaiti			2.17	
	Makohine	Tuha Sand					
	Tikapu	Hautawa Shellbed					
	Mangarere	Te Rimu Sand					
Paparangi	Mangaweka Mudstone Formation		Eagle Hill		2.7 ± 0.3	2.88	2.85 ± 0.2
			Kowhai			2.88	
Utiku	Manui		Siberia			3.12	3.12 ± 0.18

Table 1: Selected tephra-fall and/or volcaniclastic deposits and their respective stratigraphic units from the Rangitikei. Ages derived from Alloway et al. (2004), Pillans et al. (1994), Naish et al. (1996), Shane et al. (1996), Seward and Kohn (1997), Naish et al. (1998), McIntyre (2002), Pillans et al. (2005) and Grant et al. (2018).

Stop 1: Broadlands Station – Scrimmys gully

Stop 1 is located 25 km northeast of Palmerston North City and is characterised by a Plio-Pleistocene sedimentary record resting unconformably above Mesozoic basement rock of the Torlesse Composite Terrane. The main axial range, a geologically young piercement structure that has undergone a majority of its uplift over the last 1 Ma, is bound by major regional structures trending approximately NE-SW that act to deform the unconformably overlying Plio-Pleistocene succession into a sharp monocline (Beu et al. 1981; Shane 1991; Trewick & Bland 2012; Rees et al. 2018c) (Fig. 1).

Stratigraphy

The geology of the area has been documented in stratigraphic studies by Ower (1943), Rich (1959), Piyasin (1966), Carter (1972), Finlayson (1980), Marden (1984), Beanland (1995), Milner (2017) and Rees et al. (2018c). The Plio-Pleistocene sediments of the Pohangina Valley have been subdivided into the Komako, Konewa and Takapari formations, in order of decreasing age (Carter 1972). A key feature of the younger Takapari Formation is the episodic occurrence of volcaniclastic deposits that have been fluvially transported into coastal plain to shallow marine environments following voluminous volcanic eruptions originating from the TVZ. At this time the direction of fluvial transport to the area was to the south to southeast, as evidenced by the paleo-current work of Brackley (1999).

During the Early Quaternary, marine basins existed on both sides of the modern-day main axial range (Fig. 2), connected by marine seaways (Beu et al. 1981; Browne 1986, 2004; Rees et al. 2018a). Progressive retreat of the sea at c. 1.6 Ma is indicated in the Pohangina area by a change from shallow marine sedimentation to intermittent influxes of volcaniclastics within fluvio-estuarine environments. East-west drainage developed c. 1 Ma, including ponding near Woodville from which an early ancestor of the Manawatu River exited, establishing an antecedent path to the Tasman Sea across the uplifting paleo-axial range.

Castlecliffian strata preserved within the lower Pohangina Valley have been correlated to Takapari Formation of Carter (1972) in mapping by Rees et al. (2018c). Identification of the widespread c. 1 Ma Potaka Pumice (Te Punga 1952; Pillans 2017) within the Pohangina Valley succession has allowed an upper contact to be set for the Takapari Formation of Carter (1972), providing a direct link to the lower boundary of the Kaimatira Pumice Sand Formation of Fleming (1953).

The Takapari Formation, cropping out in Pohangina Valley, consists of sandy mudstones, sandstones and shellbeds interbedded with vitric volcaniclastic sandstone and carbonaceous mudstone. The lower Waitokanui Member is composed of laminated sandy mudstone with lenses of siltstone and fossiliferous horizons. This unit is unconformably overlain by the Awahou Member, comprising a succession of coarse to fine grained sandstones interbedded with massive siltstone and laminated mudstone. Pumice first appears in bed 9 and 10 of the presented section (Fig. 8), defining the base of Awahou Member. The volcaniclastic units occur in close association with high energy bed forms, such as cross bedding with erosional lower contacts, overlain by carbonaceous mudstone and common thin beds of lignite.

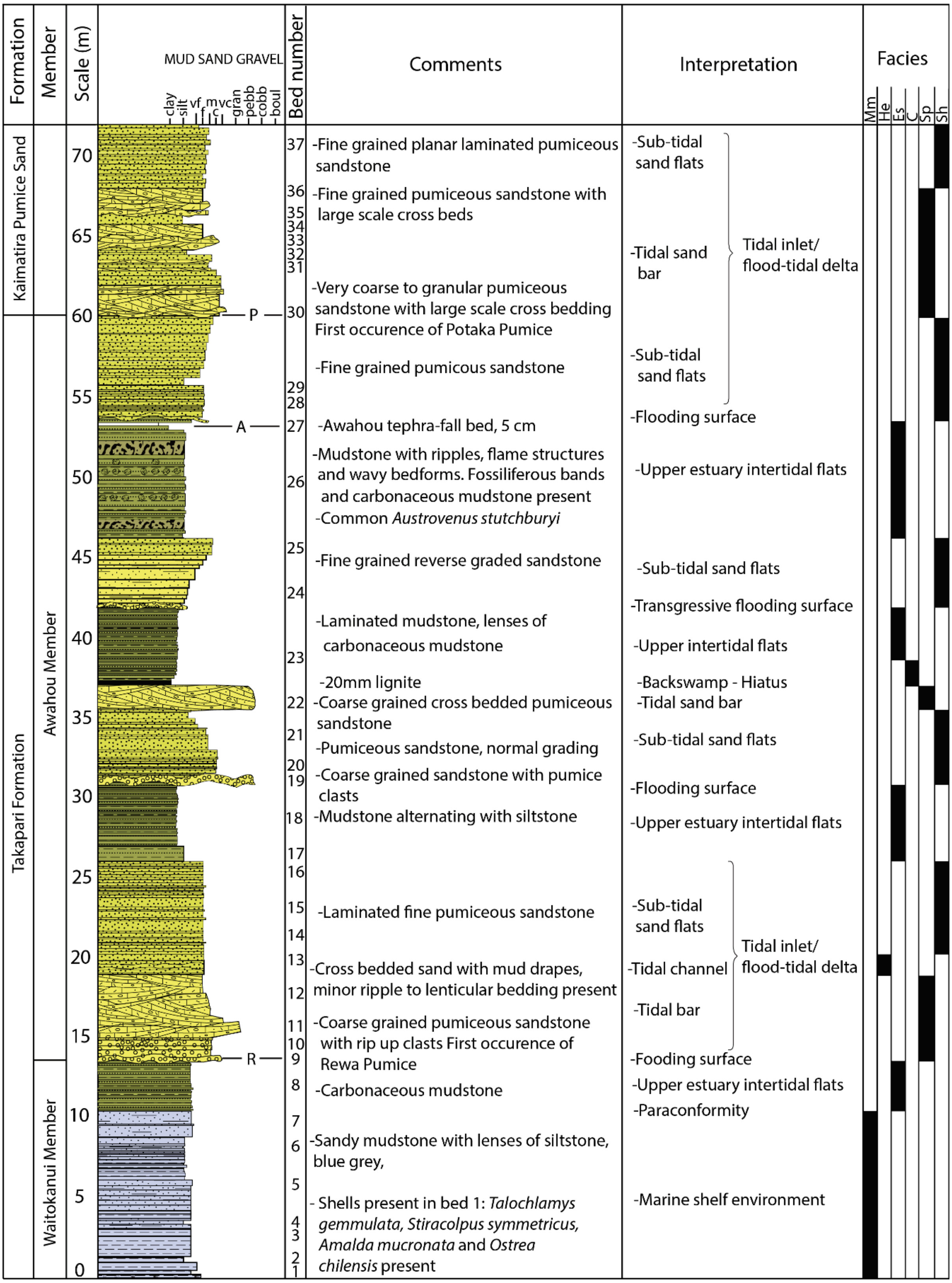

Fig. 8: Stratigraphic column of Awahou Member, Takapari Formation exposed on Awahou South Road, Broadlands Station, lower Pohangina Valley (NZGD2000 -40.280017 S, 175.773869 E).

Stop 2: Finnis Road – Pohangina Anticline

Stop 2 takes us northward up the Pohangina Valley, to a deeply incised gully system that exposes weakly consolidated to unconsolidated c. 1.2 – 1 Ma sedimentary units on the western limb of the Pohangina Anticline. The area was originally cleared in the early 20th century for pastoral farming, triggering an increase in erosion (Hancox & Wright 2005). In 1941, the Soil Conservation and River Control Act was passed to protect the area, leading to establishment of the Manawatu Catchment Board in 1944 (Miri 1999). This in turn led to establishment of the 70 ha Te Awa Farm in 1945, where erosion control trials using engineered structures, vegetation and improved livestock and pasture management techniques took place up until 1964 (Brougham & O'Connor 1979). The trials at Te Awa Farm suggested vegetation was the most cost effective and efficient method of erosion control, promoting gully retirement, space planted poplars and willows and afforestation with natives, eucalyptus, sycamores, Tasmanian blackwoods and Pinus radiata. Soil conservation work was carried out by landowners and the Catchment Board through the 1960's and became increasingly common by 1967 with establishment of the Pohangina-Oroua Catchment Control Scheme.

Gully systems draining the Pohangina Anticline such as Goulters Gully (Fig. 11 and 12) have been targeted for retirement and plantation forestry in an effort to reduce erosion. Goulters gully was initiated following forest clearance around 1900 on what appeared to be flat, fertile farmland. By 1950 a gully system approximately 11km long and 80m deep had established where there had initially been flat farmland (Galloway 2017). Willow and poplar planting was undertaken from 1967 and the catchment board planted approximately 70,000 pine seedlings between 1972-1978.

Fig. 11: Goulters Gully in 1986 from Galloway (2017).

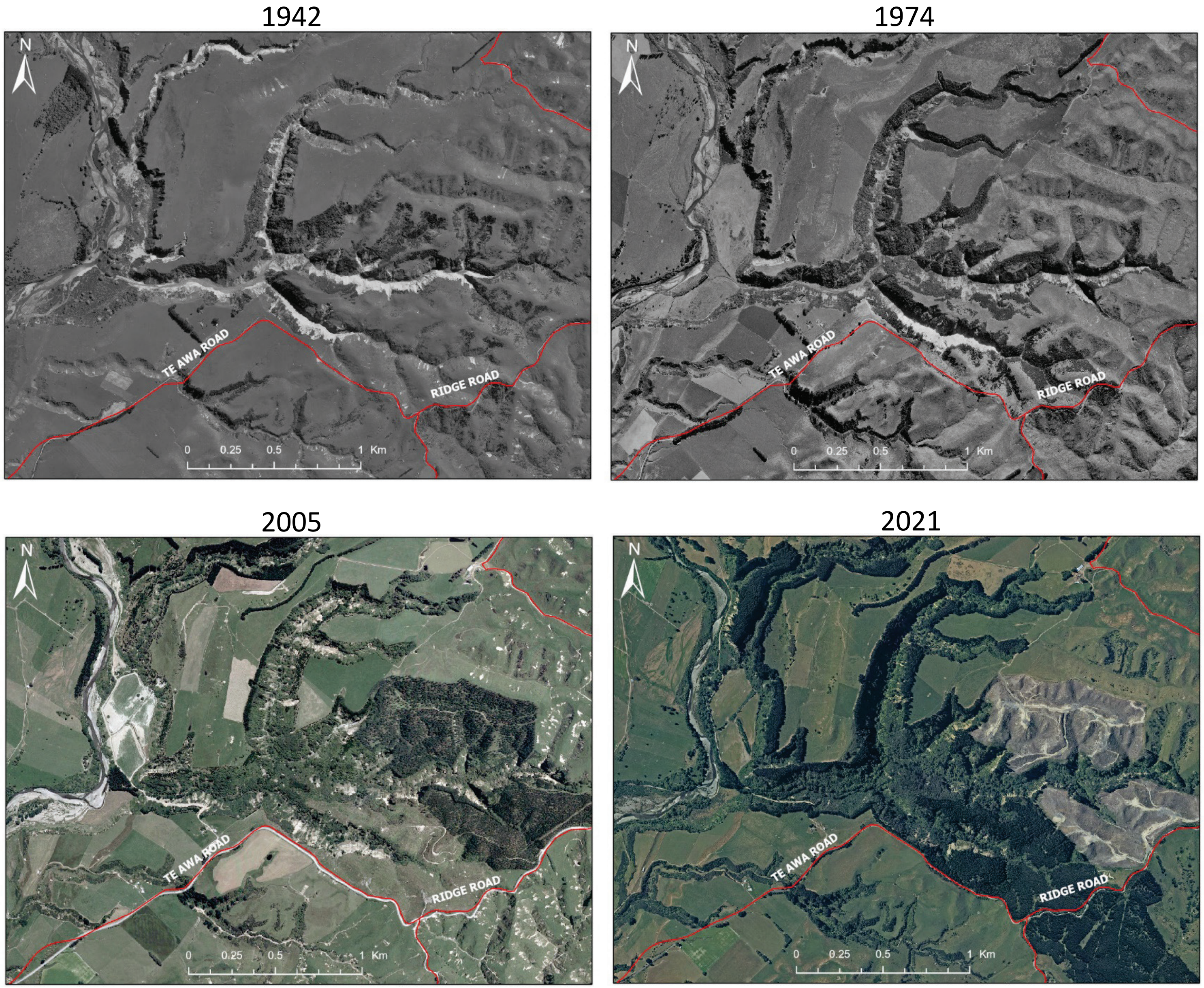

Fig. 12: Aerial photographs of Goulters Gully. Note planting in 1974, extensive soil slip in 2005 following the February 2004 storm and some harvesting in 2021.

Pohangina township is built on a fan of material that has eroded out of the weakly to unconsolidated sediments of the Pohangina Anticline (Fig. 13). In February 2004, a large storm event impacted the area causing flooding and erosion (Fuller 2005). Streams flowing out over the fan surface could be prone to future avulsion, particularly if sediment supply from erosion farther upstream allows them to build up their beds and sit high in the landscape. Evidence of gully erosion can be observed from Finnis Road, note the plantation forestry has recently been harvested in the area making it easier to observe some of the gully systems on the lower valley floor.

Stratigraphy

The Rewa Pumice (c. 1.2 Ma) occurs at the base of the Finnis Road section in parallel bedded pumiceous sand with occasional cross bed sets. There is approximately 80m of sediment between the base of the section and the Potaka Pumice (c. 1 Ma) near the top of the section. Between these two pumiceous rich units are a range of predominantly sandy units with occasional silt and mud lenses and scattered greywacke pebbles. Sedimentary structures are dominated by parallel beds but also include flaser, lenticular and convolute beds, ripples, cross bedding, rip up clasts and small channels (up to 1m wide and 15cm deep with common rip up clasts). The top of the section is characterised by a sudden influx of pumiceous rich material and contains a variety of sandy units with common cross bedding and small pebble to granule sized, sub-rounded clasts of pumice. This interval of predominantly parallel bedded, sandy sediments been the Rewa and Potaka pumices has been interpreted to represent shallow marine deposition, possibly in a nearshore to beach type environment (Brackley 1999). This is consistent with the occurrence of shellbeds just below and above the Rewa Pumice in nearby sections (Townsend 1993). The occurrence of lignite beds and rootlets in the nearby Beehive Creek section provides some evidence of terrestrial conditions; these deposits are particularly common following deposition of the Potaka Pumice at c. 1 Ma.

Sedimentary response to ignimbrite emplacement involves inundation of river and coastal settings, causing valley aggradation, river avulsion, infilling of coastal embayment's and coastal progradation (Manville & Wilson 2004; Kataoka 2011). A dramatic flood of volcaniclastic sediment entered the WB at c. 1 Ma (Potaka Pumice of Te Punga, 1952). This influx of volcaniclastic sediment forms one of the thickest transgressive system tracts (TST) documented within the WB (Pillans et al. 2005) and resulted in coastal progradation (Lewis 2007). Common convolute bedding during this interval has been suggested to be a result of increased sedimentation characterised by an abundance of unstable, water saturated sediment entering the basin, creating a hydraulically active seafloor environment (Abbott & Carter 1999).

Several authors suggest the presence of a nearby fluvial outlet supplying abundant volcaniclastic and siliciclastic sediment to the marginal marine to shallow marine environments (Townsend 1993; Brackley 1999; Rees et al. 2019). Evidence of possible fluvial reworking includes channel structures and rip up clasts of siltstone and lignite (Brackley 1999). East of the Rangitīkei, iron oxide cemented, clast supported conglomerates of sub-rounded greywacke pebbles to cobbles begin to occur around the Potaka and Kaukatea pumices (c. 0.9 – 1 Ma). The conglomerate units contain predominantly equant shaped clasts and have a distinct lack of any bioclastic material. West of the Rangitikei-Oroua interfluve the equivalent conglomerate units contain abundant crushed shell material. We will observe one such unit at our last stop in the Waitapu Stream valley.

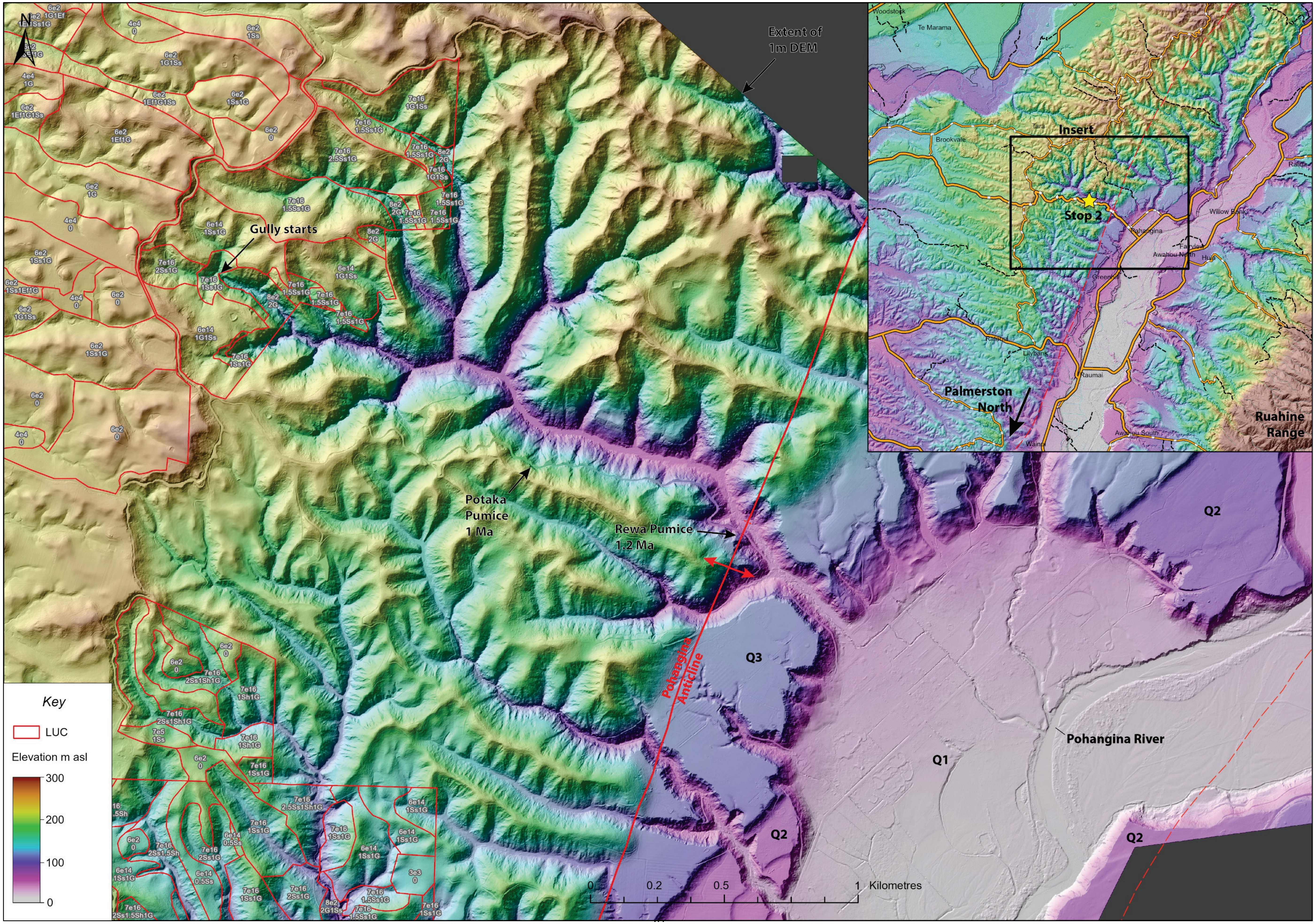

Fig. 13: Map of the area surrounding Finnis Road (Stop 2). The 1m DEM is from Horizons Regional Council.

Anticlines

The Manawatu plains contain a series of approximately NE-SW trending anticlines, including the Pohangina, Feilding, Marton, Mt Stewart-Halcombe, and Oroua anticlines (Fig. 13). These anticlines were first documented in published work by Feldmeyer et al. (1943) and Te Punga (1952, 1957). They are bounded on their eastern flank by westward dipping reverse faults that only rarely reach the surface (Begg et al. 2005) and have been identified by seismic reflection surveys (Feldmeyer et al. 1943; Anderton 1981; Melhuish et al. 1996; Jackson et al. 1998; Nicol & Beavan 2003; Lamarche et al. 2005). The anticlines are asymmetric, characterised by gentle western slopes (typically 2-5°) and steep eastern slopes (up to 70°).

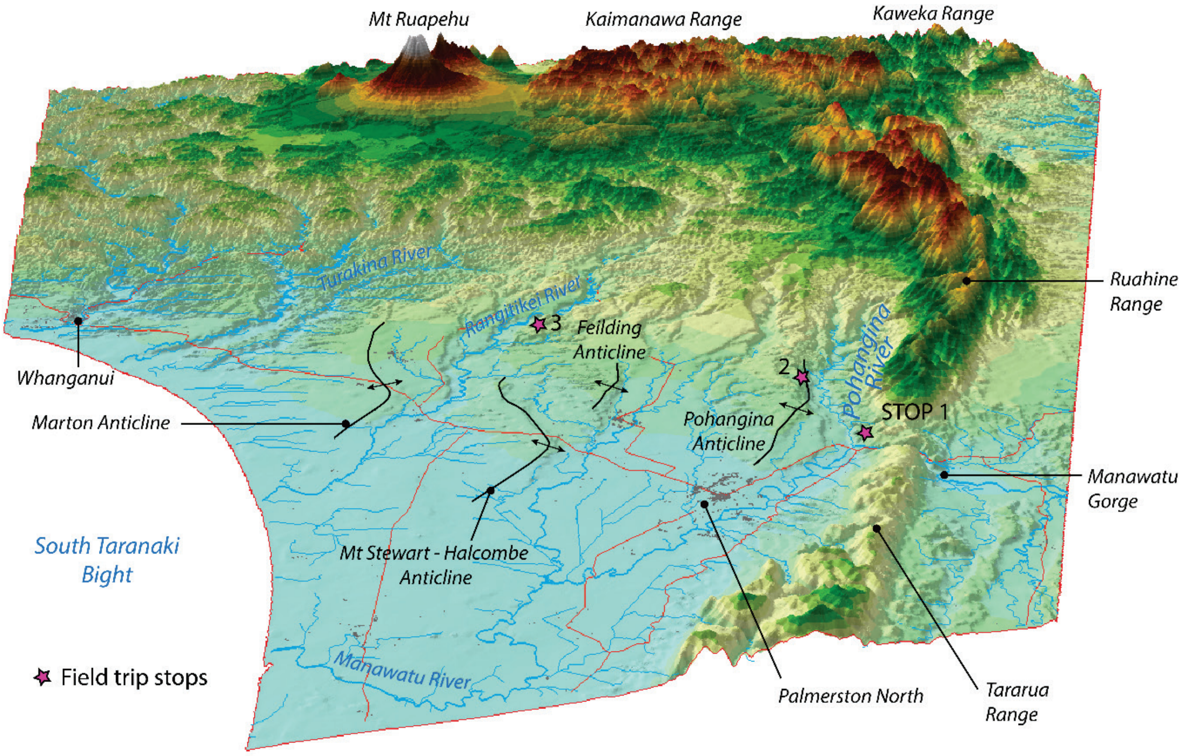

Fig. 14: 3D image of the Whanganui-Manawatu area showing major rivers and anticlines across the Manawatu Plains. DEM sourced from LINZ, 2016.

Their asymmetric geometry is interpreted to reflect the westward dipping, shallow reverse faults beneath (Jackson et al. 1998). Feldmeyer et al. (1943) and Te Punga (1957) both suggested that these anticlines were formed by ridges of basement rock trending approximately parallel with the main axial range. Structural interpretation of the Whanganui-Manawatu area indicates that these folds are still actively growing (Jackli 1957; Te Punga 1957; Lensen 1977). Preservation of the 310 ka Brunswick marine terrace tread (Pillans 1983; Pillans 2017) on the anticlines together with seismic reflection evidence of faulting within underlying Plio-Pleistocene sediments (Melhuish et al. 1996) suggests these features are long lived and have been growing since at least Pliocene time with a majority of the present day topography forming over the last c. 300 ka.

Stop 3: Waitapu Stream – Rangitikei Valley

Stop 2 is located 37 km northwest of Palmerston North City and is characterised by an important Quaternary sedimentary record, including the type section for the Waitapu Shell Conglomerate (c. 0.9 Ma). The Waitapu Shell Conglomerate is an important marker horizon in the eastern Whanganui Basin, occurring within a Pleistocene volcaniclastic record that contains early eruption products from the Taupo Volcanic Zone. The unit comprises a cross-bedded pebbly-shell conglomerate containing the first influx of Kaukatea Pumice (c. 0.9 Ma) within the Rangitikei succession.

The fossil assemblage contains a mixture of estuarine to offshore species including a component derived from underlying rocks. Deposition is thought to have taken place seaward of a transgressive shoreline on a storm-dominated, muddy, innermost shelf, characterised by migrating gravel dunes interfingering with deeper water heterolithic facies. The abundance of fresh pumice and charcoal indicates deposition closely followed a major TVZ eruption within a phase of renewed volcanic activity from the Mangakino Volcanic Centre at c. 0.9 Ma.

Waitapu Stream – Type section

The Waitapu Shell Conglomerate (Fig. 15) contains a heterolithic mixture of bioclastic and siliciclastic sediments, with clasts typically of pebble size in a matrix of coarse sands and granules. The basal contact is erosive with channels up to 1 m deep carved into the underlying laminated to lenticular bedded siltstone and fine sandstone. Large-scale, low-angle, trough cross-bedded sets 1-2 m thick contain large (up to 400 mm), angular rip-up clasts of the laminated to lenticular bedded siltstone and fine sandstone. Lenses of siltstone, 0.1 – 0.3 m thick, drape the crossbeds, pinching out laterally. The rip-up clasts contain original laminated to lenticular bedding formed prior to reworking and appear to be derived from the heterolithic beds immediately beneath the cross-bedded sets.

Clasts are dominantly pebbles of sub-rounded, disc to blade-shaped, greywacke, pumice, rhyolite and ignimbrite 2–64 mm in diameter, with rare larger clasts >100 mm. Clasts are typically weakly aligned along the cross-bed sets. Rare charcoal is particularly evident within the lower-most channelled cross-bed sets, occurring as pockets of rounded clasts up to 70 mm in diameter. Variable concentration and composition of clasts occurs throughout the deposit, with particular cross-bedded sets enriched in bioclastic material, greywacke pebbles and pumiceous alluvium. Lenses of distinctive, very coarse to granular quartz- and ferromagnesian-rich sands occur locally, displaying strong sorting and resultant mineral separation by density.

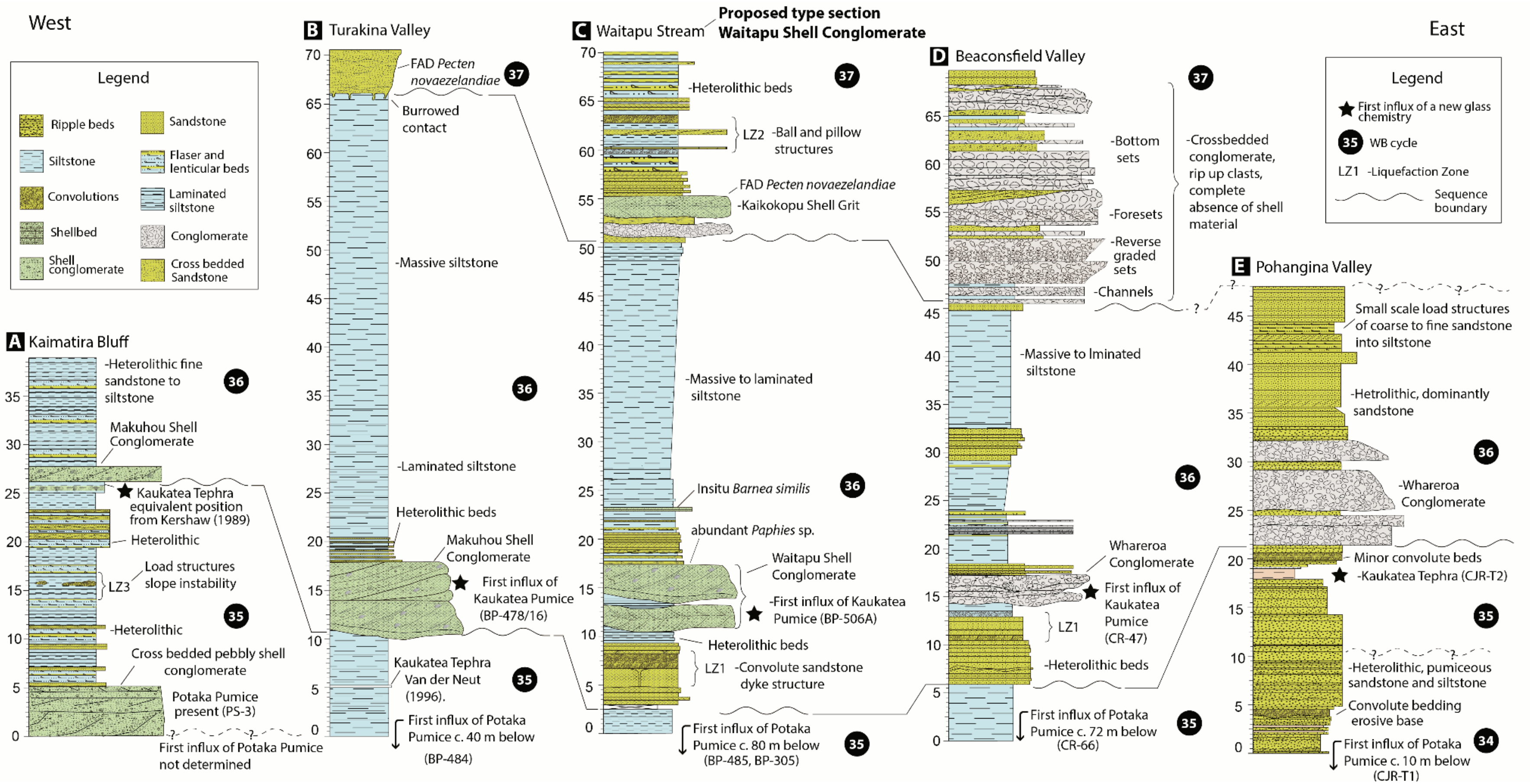

Fig 15: Stratigraphic log of the type section for the Waitapu Shell Conglomerate near the mouth of Waitapu Stream, Rangitikei Valley (BL34 1582 6596) and other laterally equivalent sections from across the basin. Data sourced from Kershaw (1989), Abbott (1994), Shane (1994), van der Neut (1996), Pillans et al. (2005), Rees et al. (2018c) and Rees et al. (2019).

Bioclastic material is highly worn, comprising approximately equal proportions of granule-sized debris and variably preserved fossil Mollusca dominated by Paphies sp. Particular lenses of highly enriched bioclastic material occur, in which shell fragments and whole shells comprise up to 80% of the material in a siliciclastic matrix. Similarly, other lenses are enriched in pumiceous sediment containing abundant sub-rounded pumice clasts up to 150 mm in diameter. The largest pumice clasts significantly outsize clasts of any other lithology.

A gradual fining upward sequence occurs above the Waitapu Shell Conglomerate, changing from heterolithic sediments with rare shell debris to massive, sparsely fossiliferous siltstone. Overlying the fine-grained massive siltstone is a conglomeratic unit. The base is characterised by a 1 m thick, well sorted, un-fossiliferous, clast-supported greywacke conglomerate overlain by coarse to granule sandstone. A cross-bedded shell conglomerate, locally scours into this greywacke conglomerate-sandstone unit. The first appearance of *Pecten novaezelandiae* within the shell conglomerate allows correlation to the Kaikokopu Shell Grit (Abbott 1994). The Kaikokopu Shell Grit passes upwards into a complex of heterolithic, interbedded sandstone, siltstone and 0.2–0.5 m thick beds of coarse shell debris and sandstone displaying unidirectional cross-bedding.

Two distinctive beds of soft sediment deformation, 0.3 and 0.6 m thick, occur within a highly variable succession of alternating lenticular fine sandstone and siltstone with rare, thin lenses of shell debris. These beds are characterised by ball and pillow structures comprising fine sandstone to siltstone tightly folded into concentric layers with some lateral variation in grain-size and patterns of deformation. The heterolithic units gradually fine upward with rare 10–50 mm coarse to granule sandstone layers present, until near the top of the succession where structureless siltstone dominates.

Macrofossil paleoecology of the Waitapu Shell Conglomerate

The fossil fauna of the Waitapu Shell Conglomerate (Fig. 16, Table 2) provide a glimpse into a range of ecological zones present at the time of deposition. Nevertheless, these interpretations must be treated with caution, given the substantial reworking, selective sorting and concentration of more robust specimens within the interpreted shallow-water high-energy environment (Cooper et al. 2006).

Some specimens, such as the single large shell of the restricted Nukumaruan species *Struthiolaria frazeri*, have been derived from underlying rocks, so a variety of other specimens likely have as well. This is not surprising, given the erosional nature of the deposit. At the Castlecliff Type Section *Barnea similis* is typically found boring the unconformable surfaces between successive cycles (Abbott & Carter 1994; Abbott & Carter 1997; Abbott 1998; Beu 2006), a relationship that also occurs in the Rangitikei area (Abbott 1994). The author has not found deposits containing *B. similis* directly below the Waitapu Shell Conglomerate. However, specimens in growth position occur within siltstone stratigraphically above the Waitapu Shell Conglomerate (Abbott 1994). We suggest rare, dislodged specimens became incorporated within the Waitapu Shell Conglomerate from laterally adjacent hard ground areas, where suitable habitat for boring bivalves was present. This hard ground likely comprised a shelf of consolidated Pleistocene siltstone exposed by wave erosion/transgressive ravinement, similar to environments along the southern Taranaki Bight today. Other species certainly have been transported far from their life environments, particularly the soft-bottom shelf-dwelling Gastropoda *Alcithoe fusus, Alcithoe arabica, Aeneator marshalli, Austrofusus glans, Amalda novaezelandiae, Amalda mucronata* and *Poirieria zelandica.*

Estuarine species *Austrovenus stutchburyi, Cyclomactra tristis, Paphies australis, Cominella glandiformis* and *Cominella adspersa* are rare in the outcrop and typically their shells are extremely abraded, suggesting substantial transportation and reworking. Dominant open coast to sandy nearshore species, particularly *Paphies donacina* and *P. delta,* while disarticulated, appear the least worn and significantly out-number other species within the assemblage, suggesting the depositional environment was supplied by a nearby high-energy shoreface environment or that reworking during transgressive ravinement principally involved erosion of shoreface deposits.

Austrovenus stutchburyi is encountered in many sand and mud flats in the outer parts of modern New Zealand estuaries, although larger shells are typically found in the outer parts of bays with higher salinity and more diverse faunas (Beu & Maxwell 1990; Morley & Hayward 2009). The genus is unknown from fully marine situations. Its presence within the Waitapu Shell Conglomerate together with a variety of other species is interpreted to represent transportation and reworking of shells from a range of environments, concentrated within a shallow marine environment by wave and tidal action. It is possible the shells of *A. stutchburyi* have been transported a considerable distance from their living habitat, carried by swift tidal currents to a marine setting (Hayward & Stilwell 1995) or reworked from underlying deposits (Abbott 1994).

C. tristis is a deep-burrowing species typically found living together with *Macomona liliana* in the same intertidal flats as *A. stutchburyi,* although located deeper in the substrate (Beu & Maxwell 1990). *Alcithoe arabica* together with *Cominella adspersa* likely actively preyed on *A. stutchburyi* within the intertidal zone, taking shelter amongst the rocky, hard ground areas inhabited by *Ostrea chilensis* and *Maoricrypta profunda* to retreat whilst laying eggs (Hayward et al. 2001; Hayward et al. 2002). *Paphies australis* is also estuarine, mainly inhabiting the tidal race channels near the mouths of estuaries, but also living near the heads of large bays (Grant-Taylor & Beu 1974; Beu & Maxwell 1990; Morley et al. 2006).

Nearshore to open coast dwelling species are abundant within the Waitapu Shell Conglomerate, including the typical modern day associations of the tuatua species *Paphies donacina* and *Paphies delta* (Beu & De Rooij-Schuiling 1982). Likely the sandy foreshore was similar to stretches of the modern-day Rangitikei coastline where tuatua live buried in the intertidal to subtidal sands. The widely ranging suspension feeding turritellid *Maoricolpus roseus* likely inhabited the sandy coastal area together with *Pelicaria vermis*, lying buried immediately beneath the surface of the sand whilst feeding (Morton 1951; Eagle & Hayward 1993; Donald & Spencer 2015). The carnivorous epifaunal gastropod *Tanea zelandica* roamed the sandy coastal area preying on burrowing bivalves.

A hard-ground environment is indicated by *Ostrea chilensis* (Beu 2010), formed by influxes of siliciclastic gravels into the shallow marine environment forming gravel dunes and perhaps localised gravelly foreshores, through wave action, acting to sort and concentrate coarse material in near shore to inner-most shelf environments whilst winnowing away fine silts and sand. Strong tidal currents allowed for siliciclastic bypass and stratigraphic condensation in areas of the sea floor, resulting in a relative enrichment of bioclastic material and creation of low angle shell banks consisting of broken and abraded shell material, crushed by wave action. In places, hard cohesive siltstone was exposed by storm wave erosion including local scouring and creation of rip up clasts, supplementing the abundance of fluvially derived greywacke clasts within hardground areas. Coastal platforms formed of consolidated Pleistocene sedimentary rock stripped bare of surficial sediment provided suitable habitat for *Barnea similis* to bore into, analogous to coastal settings along the southern Taranaki coastline today.

Evidence of a relatively stable gravelly area composed of a mixture of broken shells, greywacke pebbles and clasts of eroded Pleistocene sedimentary rock is suggested by the presence of *Barbatia novaezelandiae*, which lives basally attached in crevices and under stones together with *Ostrea chilensis and Chamaesipho columna* (Hayward & Morley 2002; Morley et al. 2006). The gastropods *Calliostoma punctulatum* and *Coelotrochus tiaratus* prefer to graze hard ground areas, suggesting that an array of bryozoans, sponges and brown algae lined the stones and ledges of the gravelly area. *Ostrea chilensis* lived here, preyed upon by the deadly oyster borer *Haustrum scobina*, which took shelter within crevices offering protection from predators, whilst *Paratrophon sp.* roamed the environment preying on other gastropods, bivalves and barnacles. Among the scavenging whelks, there are two species that lived in this zone including *C. nassoides elegantula* and *Buccinulum vittatum.* We interpret few bivalves to have lived within this gravelly zone, although it was likely inhabited by mussels, oysters and rock borers.

More widely ranging species such as *Purpurocardia purpurata* and *Penion sulcatum* inhabited a soft-bottom zone at inner shelf depths together with the soft-bottom gastropods *Alcithoe sp*. *Aeneator marshalli*, *Austrofusus glans*, and *Poirieria zelandica* (Hayward et al. 1999; Beu 2011). Carnivorous gastropods roamed the soft ground zone such as *Alcithoe arabica,* able to move with ease on soft substrates, feeding on bivalves and laying their eggs in shelly or stony zones of the sea floor (Ponder 1970). Other common gastropods occupying the soft-ground area include the olive shells *Amalda mucronata* and *A. novaezelandiae.*

Waitapu Shell Conglomerate – Depositional Environment

The Waitapu Shell Conglomerate contains fossils representative of a wide variety of depositional environments. Shells are well mixed, broken and abraded, typical of reworking by waves and tidal currents. Trough cross-bedding indicates dune bed forms produced by strong unidirectional currents (Harms et al. 1982; Duke et al. 1991). Set thicknesses of up to 2 m implies the influence of strong tidal and/or storm-driven currents (Abbott 1998; Le Bot & Trentsaux 2004). Mud draped crossbed foresets indicate fluctuating energy conditions, including slack water/quiescent periods where finer particles settled from suspension. The channelled erosion surface below the crossbed sets was likely formed by strong storm/tidal action, possibly related to a reduction in the mean wave base and localised scour and fill of sediments (Dott & Bourgeois 1982).

The Waitapu Shell Conglomerate is analogous to the basal TST shell conglomerates described at the Castlecliff section (Abbott 1998), interpreted to represent deposition on a storm-dominated, muddy innermost shelf (Abbott & Carter 1994). We infer the presence of a fluvial outlet in close proximity to the depositional site, with headwaters located in the present day central North Island, supplying abundant greywacke and volcaniclastic detritus to the coast. This is consistent with the lateral changes observed within the Waitapu Shell Conglomerate, as it grades into an eastern margin conglomerate facies, characterised by a lack of marine fossils, rare lignite and equant greywacke clasts, suggesting closer proximity to and influence from fluvial systems draining the uplifting paleo-axial range.

At the WBs eastern margin, we envision braid plains, delta fronts and gravelly beaches to have developed during sea level low stand, locally incising through the previous HST, as evidenced by delta type architecture above undulating surfaces of erosion within eastern margin conglomerate deposits (Fig. 17). Nearby land was likely deforested and sediment input particularly high (Beu et al. 1981; Lewis et al. 1994).

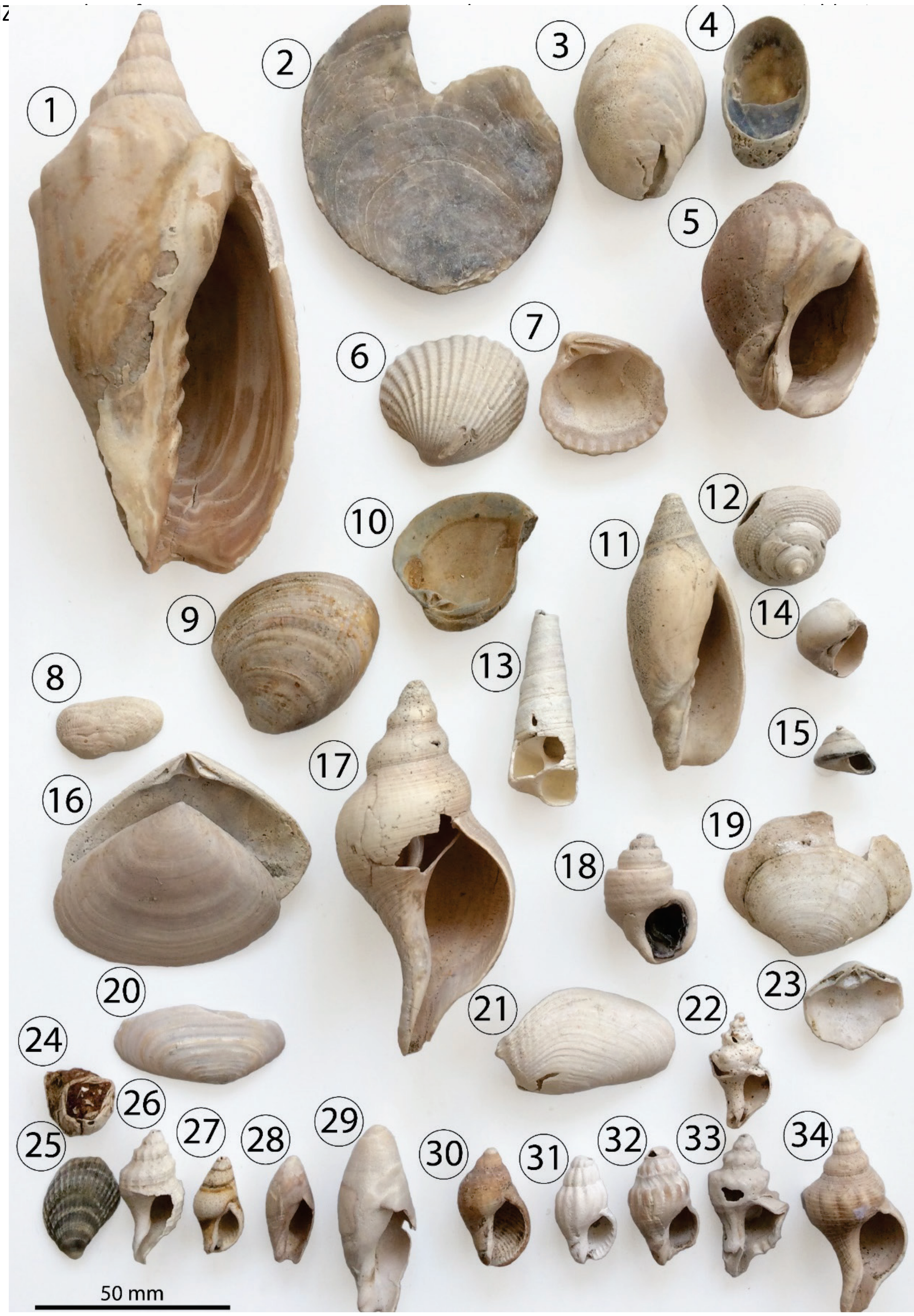

Fig. 16: Fossils from the Waitapu Shell Conglomerate that represent a variety of subtidal to offshore environments: 1) *Alcithoe arabica*; 2) *Ostrea chilensis*; 3, 4) *Maoricrypta profunda*; 5) *Cominella adspersa*; 6, 7) *Purpurocardia purpurata*; 8) *Barbatia novaezelandiae*; 9, 10) *Austrovenus stutchburyi*; 11) *Alcithoe fusus*; 12) *Calliostoma punctulata*; 13) *Maoricolpus roseus*; 14) *Tanea zelandica*; 15) *Coelotrochus tiaratus*; 16) *Paphies delta*; 17) *Aeneator marshalli*; 18) *Pelicaria vermis*; 19, 23) *Cyclomactra tristis*; 20) *Paphies australis*; 21) *Barnea (Anchomasa) similis*; 22, 33) *Poireiria zelandica*; 24) *Chamaesipho columna*; 25) *Lima zelandica*; 26) *Austrofusus glans*; 27) *Buccinulum vittatum*; 28) *Amalda novaezelandiae*; 29) *Amalda mucronata*; 30) *Haustrum scobina*; 31) *Cominella nassoides elegantula*; 32) *Cominella glandiformis*; 34) *Penion sulcatum*

Table 2: List of macrofossils collected from Waitapu Shell Conglomerate type section.

Phylum	Class	Genus and species	Environment
Mollusca	Bivalvia	*Ostrea chilensis* Philippi (Küster, 1844)	Rocky intertidal
		Barnea (Anchomasa) similis (Gray, 1835)	
		Barbatia novaezelandiae (Smith, 1915)	
		Austrovenus stutchburyi (Wood, 1828)	Estuarine/lagoonal
		Cyclomactra tristis (Reeve, 1854)	
		Paphies australis (Gmelin, 1791)	
		Paphies delta Beu, 2006	Open coast/shoreface
		Paphies donacina (Spengler, 1793)	
		Purpurocardia purpurata (Deshayes, 1854)	
		Lima zelandica (Sowerby, 1877)	Soft-bottom shelf
	Gastropoda	*Calliostoma punctulatum* (Martyn, 1784)	Rocky intertidal
		Coelotrochus tiaratus (Quoy & Gaimard, 1834)	
		Buccinulum vittatum (Quoy & Gaimard, 1833)	
		Haustrum scobina (Quoy & Gaimard, 1833)	
		Paratrophon sp.	
		Cominella adspersa (Bruguière, 1789)	Estuarine/lagoonal
		Cominella glandiformis (Reeve, 1846)	
		Maoricolpus roseus (Quoy & Gaimard, 1834)	Open coast/shoreface
		Maoricrypta profunda (Hutton, 1873)	
		Tanea zelandica (Quoy & Gaimard, 1832)	
		Pelicaria vermis (Martyn, 1784)	
		Cominella nassoides elegantula (Finlay, 1926)	Soft-bottom shallow water to shelf
		Alcithoe arabica (Gmelin, 1791)	
		Penion sulcatum (Lamarck, 1816)	
		Alcithoe fusus (Quoy & Gaimard, 1833)	Soft-bottom shelf
		Aeneator marshalli (Murdoch, 1924)	
		Austrofusus glans (Röding, 1798))	
		Amalda novaezelandiae (Sowerby, 1859)	
		Amalda mucronata (Sowerby, 1830)	
		Poirieria zelandica (Quoy & Gaimard, 1833)	
Arthropoda		*Chamaesipho columna* (Spengler, 1790)	Rocky intertidal

Low stand deposits were later scoured and eroded during transgressive ravinement, reworked into shelly gravel dunes at inner-most shelf depths (Abbott 1998). Local preservation of low stand deposits in eastern WB (Abbott 1992), suggests transgressive ravinement was either not as destructive or that low stand incision was more pervasive along the eastern basin margin. Marine transgression and associated shoreface retreat triggered sedimentary infilling of the valleys incised during the previous low stand (Rees et al. 2018b). Estuarine systems developed, hosting a wide range of species, including *Cyclomactra tristis, Austrovenus stutchburyi, Paphies australis* and *Cominella adspersa,* the shells of which were transported into the marine environment by rivers (Hayward & Stilwell 1995) or became entrained during transgressive ravinement and sedimentary reworking (Abbott 1998).

The abundant unweathered pumice within the Waitapu Shell conglomerate is typically of a much larger clast size than the accompanying clasts of greywacke, suggesting that a phase of important eruptions in the TVZ preceded and perhaps accompanied its deposition (Te Punga 1952). This is consistent with the incorporation of clasts of charcoal within the lowest beds of the Waitapu Shell Conglomerate. In the Pohangina, Turakina and Whanganui valleys, Kaukatea Tephra is preserved within marginal to shallow marine deposits just below the stratigraphic position of the Waitapu Shell Conglomerate, suggesting that volcanic activity closely preceded deposition of Waitapu Shell Conglomerate, occurring within cycle 35 (Naish et al. 2005; Pillans et al. 2005).

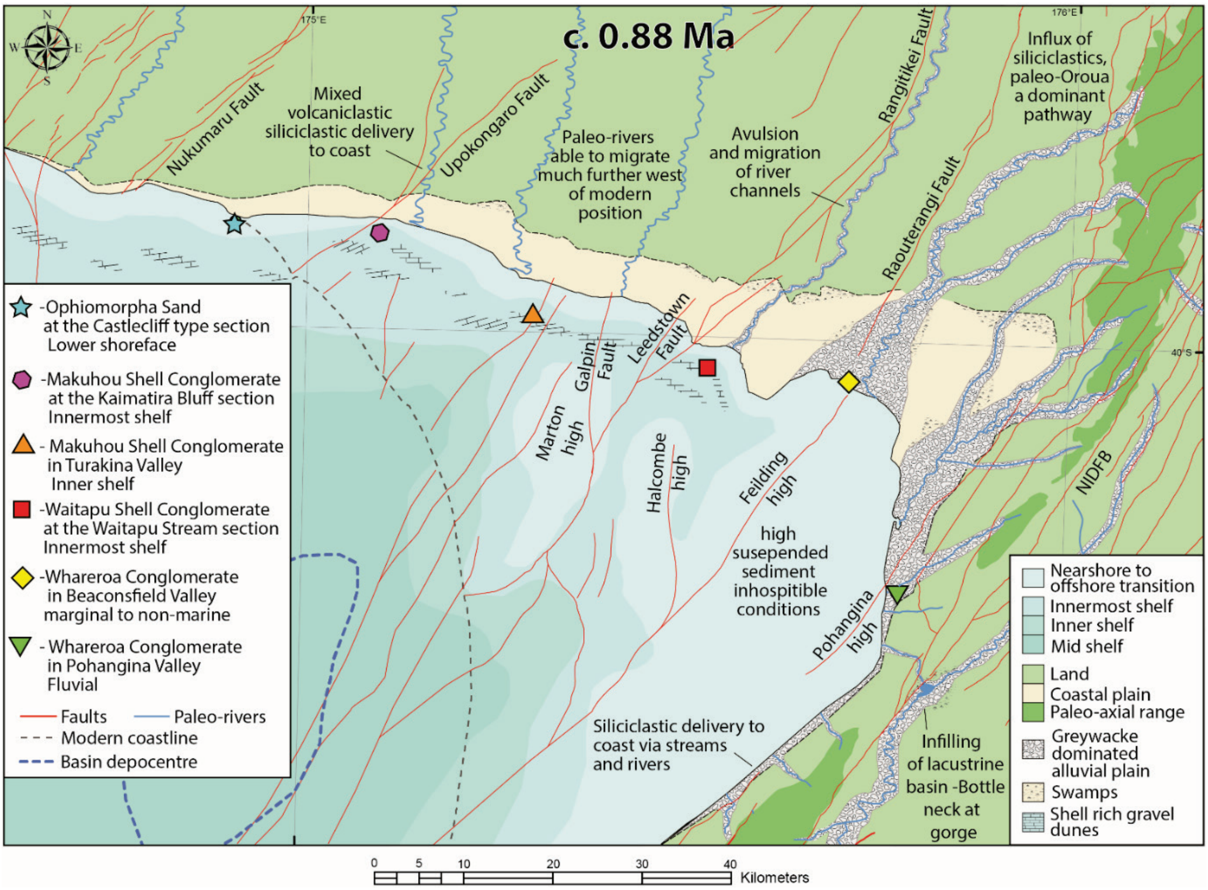

Fig 17: Schematic cartoon of the WB during the marine transgression of WB cycle 36 (MIS 22) at c. 0.88 Ma (Bowen et al. 1998; Lisiecki & Raymo 2005).

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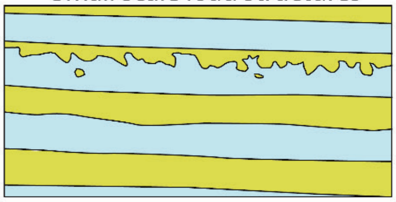
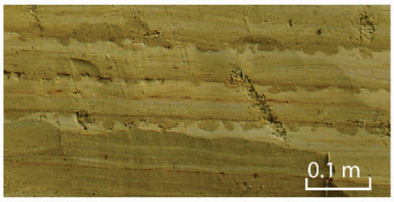
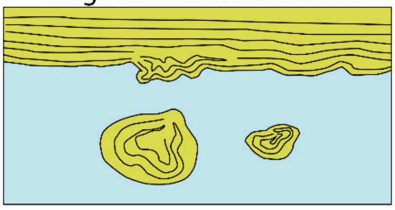
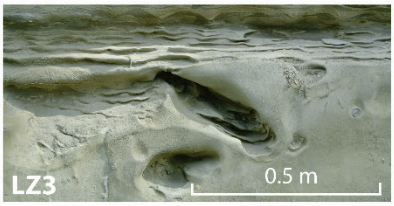
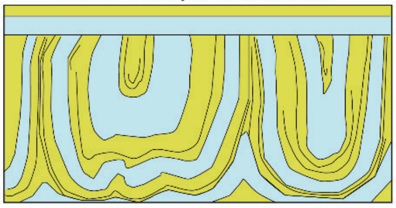
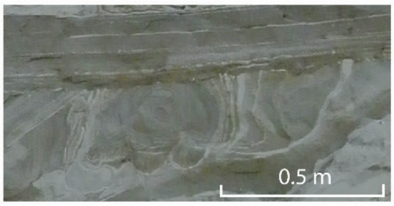
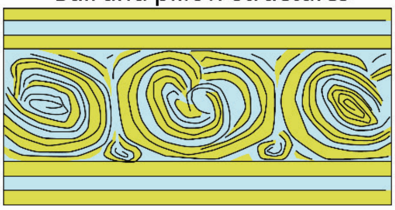
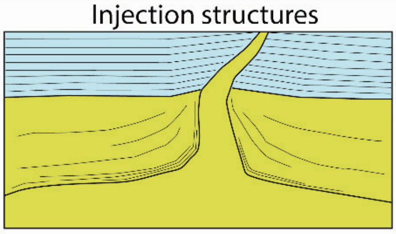
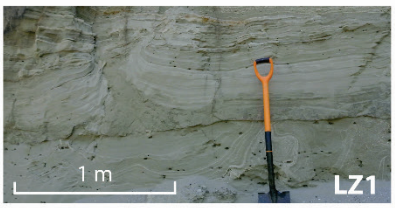
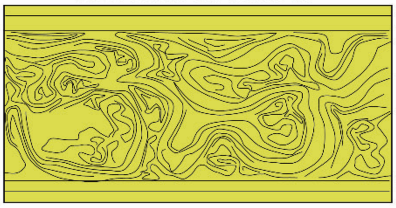
Appendices

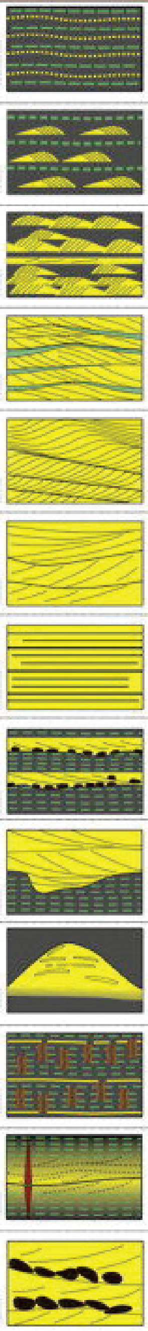
Sedimentary structures you might see on the field trip

<table>
<tr><th></th><th>SSDS type</th><th>Field example</th><th>Deformation processes</th></tr>
<tr><td>A</td><td>Small scale load structures
</td><td></td><td rowspan="4">Gravitationally unstable reverse density gradient involving sands over silts

Commonly truncated along upper contact suggesting deformation occured at or close to the sediment/water interface

Large scale load structures are associated with evidence of slope instability along the margins of muddy swales between gravelly dunes and channels</td></tr>
<tr><td>B</td><td>Large scale load structures
</td><td></td></tr>
<tr><td>C</td><td>Water escape structures
</td><td></td></tr>
<tr><td>D</td><td>Ball and pillow structures
</td><td></td></tr>
<tr><td>E</td><td>Injection structures
</td><td></td><td rowspan="2">Liquefaction and fluidisation involving injection of sands

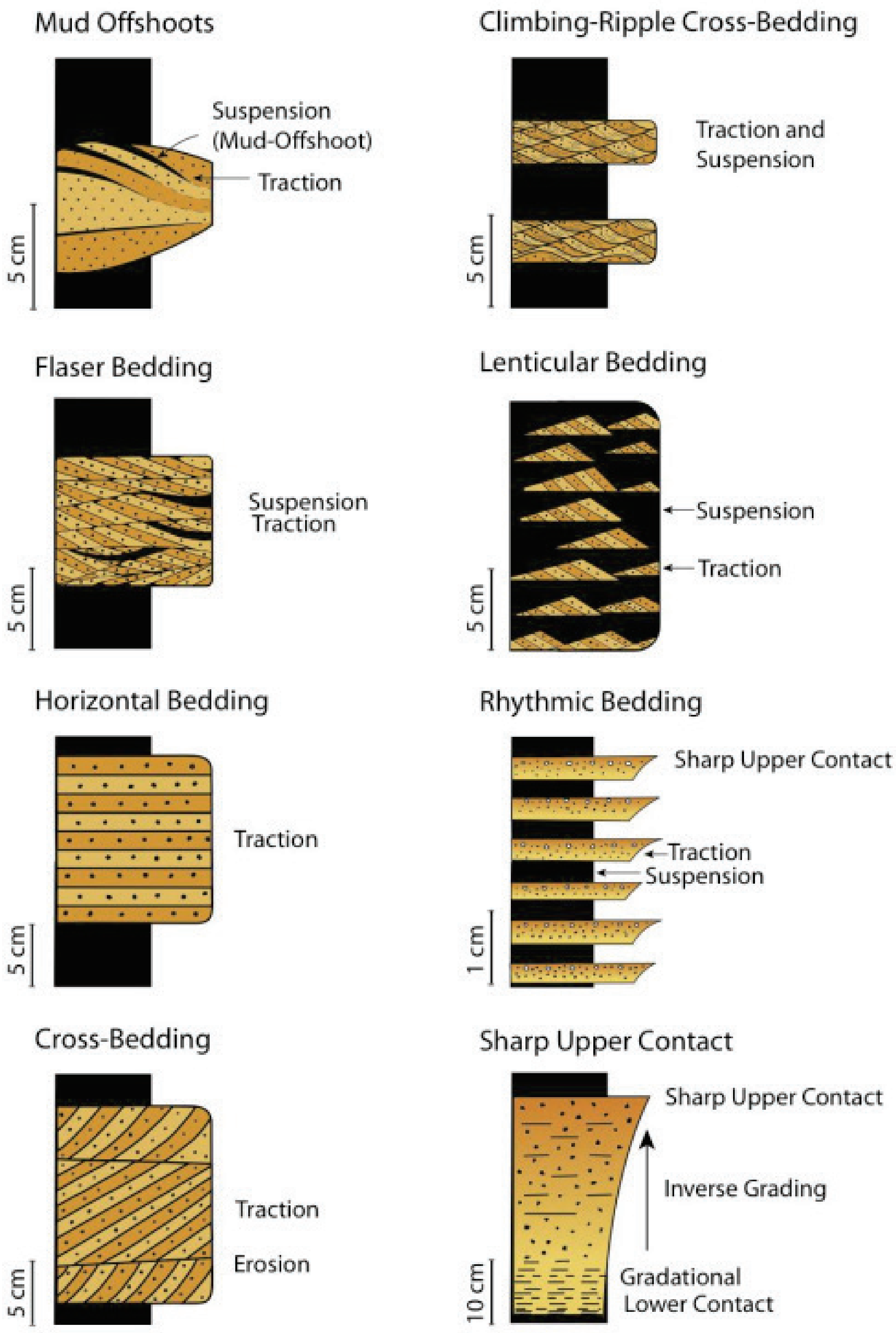
Lower beds form massive to indistinct bedding following redistribution and settling of grains</td></tr>
<tr><td>F</td><td>Convolute structures
</td><td></td></tr>
</table>



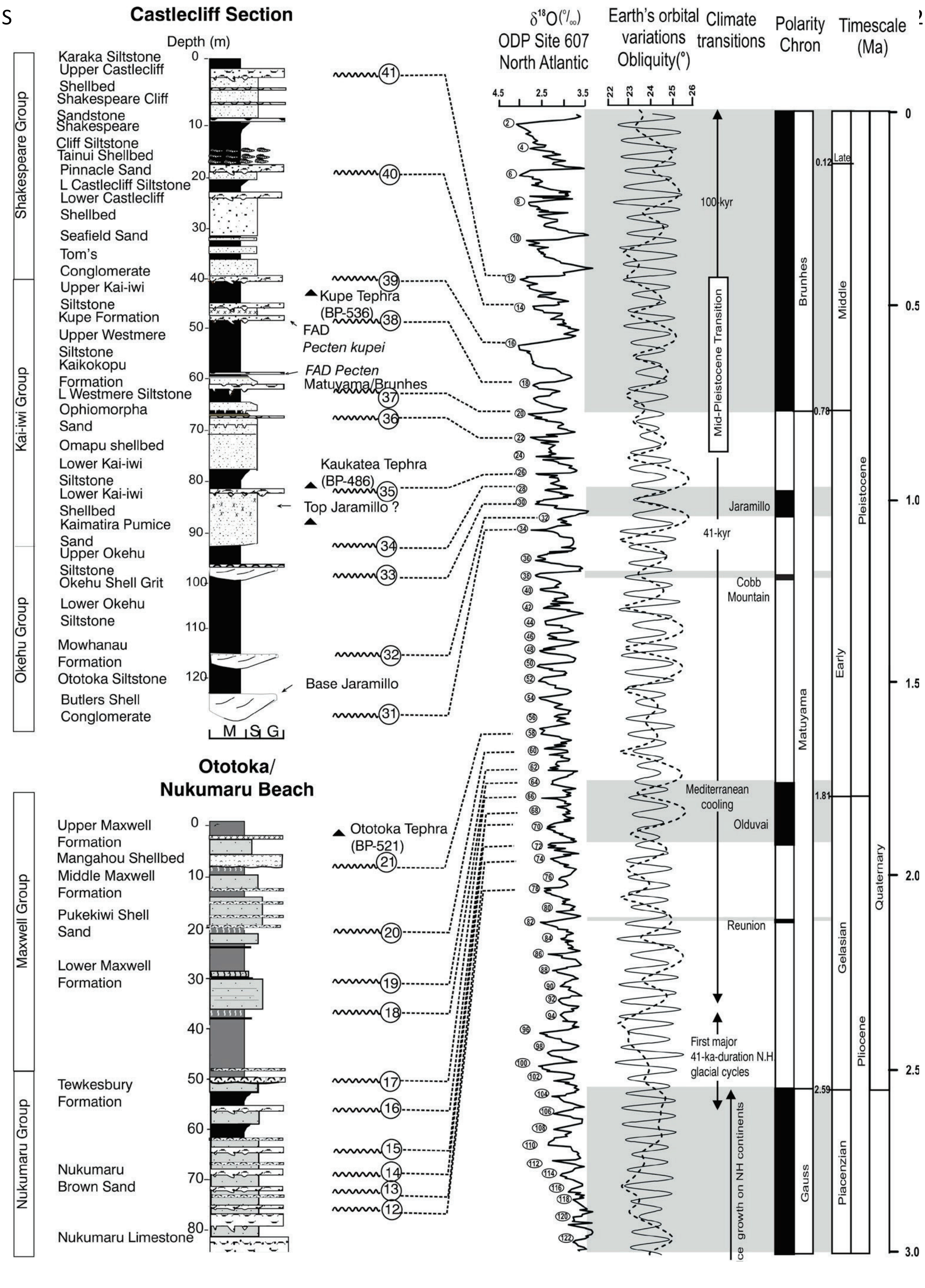
Sketch	Sed. structures	Dominant grain size	Enviromental implications
1 cm	Horizontal or sinusoidal lamination, stripped, fine-grained deposits; "wispy" lamination	Fine sand, silt & mud < 2 Ø < 0.250 mm	Low current strength Predominance of deposition from suspension
1 cm	Lenticular bedding starved ripples	Fine sand, silt & mud < 2 Ø < 0.250 mm	Alternating flow conditions, low to moderate current strength, winnowing
1 cm	Wavy bedding, flaggy chalks	Fine sand, silt & mud < 2 Ø < 0.250 mm	Alternating flow conditions, low to moderate current strength
1 - 5 cm	Flaser bedding, mud offshoots	Fine sand to silt 8 - 2 Ø 0.004 - 0.250 mm	Alternating flow conditions, Current speed = 10 - 40 cm / s
1 - 5 cm	Climbing ripples (subcritical to supercritical)	Very fine to medium sands 4 - 1 Ø 0.063 - 0.5 mm	Current speed = 10 - 40 cm / s High suspension load
10 - 50 cm	Large-scale cross-bedding, megaripples, dunes, sandwaves	Medium sands 2 - 1 Ø 0.250 - 0.5 mm	Current speed = 40 - 200 cm / s Barchan dunes usually form at 40-80 cm / s
1 cm	Parallel lamination (upper stage plane beds), presence of primary current lamination	Very fine to medium sands 4 - 1 Ø 0.063 - 0.5 mm	Current speed = 40 - 200 cm / s
1 cm	Minor erosive surfaces, mud rip-up clasts, upper sharp contacts	Sand, silt & mud < -1 Ø < 2 mm	Alternating flow conditions, low to moderate current strength
1 - 5 cm	Sole marks: flutes, obstacle scours & longitudinal scours, cut & fill structures	Sand, silt & mud < -1 Ø < 2 mm	Flow speed peaks
5 cm	Longitudinal ripples	Coarse sandy muds (20 % sand)	Low current speed = 2 - 5 cm / s Winnowing
1 - 10 cm	Bioturbation (strongly variable)	Sand, silt & mud	Low current speed Strong paleoecological control, Low to moderate accumulation rates
3 - 20 cm	Normal & reverse grading at different scales and within different types of deposits	From coarse sand to mud Usaully fine sand, silt & mud	Gradual changes in flow strength
0.1 - 2 cm	Pebble lags, furrows	Coarse sand, microconglomerate	Current speed over 200 cm / s

Clay and / or mud · Silt · Sand · Pebbles & cobbles

Different types of sedimentary structures from Rebesco et al. (2014)

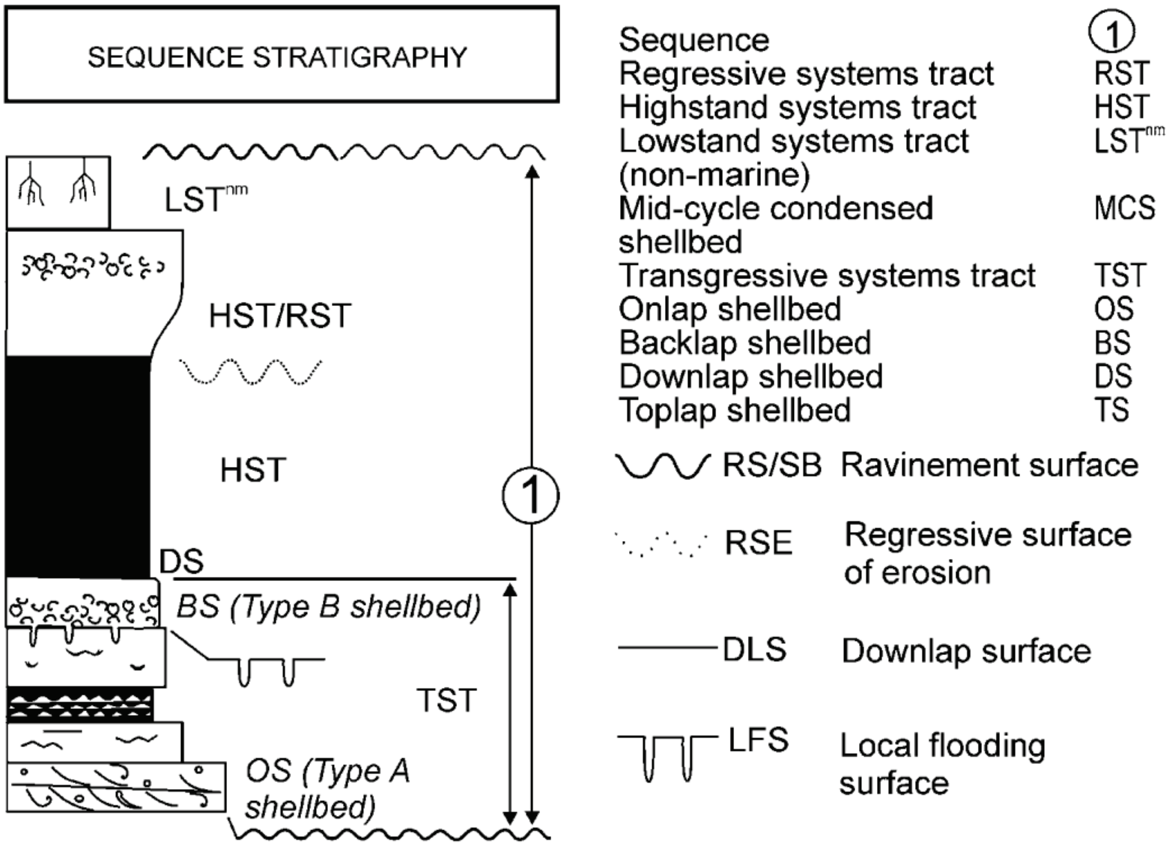


Different types of sedimentary structures from Rebesco et al. (2014)



Composite stratigraphic columns for the Nukumaru and Castlecliff coastal sections showing lithostratigraphy, sequence stratigraphy, magnetostratigraphy and correlations with the oxygen isotope timescale from Naish et al. (2005)

Sequence Stratigraphy Terminology



Sequence stratigraphic terminology from Abbott et al. (2005)