Waipaoa Sedimentary System Fieldtrip

Integration and Synthesis of MARGINS Sediment Source-to-Sink Research Workshop

Gisborne, April 2009

Field trip leaders:
Nicola Litchfield
Eric Bilderback
Jesse McNinch
Michael Marden
Matthew Wolinsky

Photo: L. Homer CN-42225

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Nicola Litchfield, GNS Science, Lower Hutt, New Zealand
Eric Bilderback, University of Canterbury, Christchurch, New Zealand
Jesse McNinch, US Army Corps of Engineers, Kitty Hawk, USA
Michael Marden, Landcare Research, Gisborne, New Zealand
Matthew Wolinsky, Shell International Exploration and Production, Houston, USA

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Introduction and setting

The main purpose of this fieldtrip is to provide an overview of the nearshore and terrestrial portions of the Waipaoa Sedimentary System (Waipaoa SS). The trip commences at Kaiti Hill lookout immediately adjacent to Gisborne City, then travels north along the Waipaoa River mainstem, before diverging along the Mangatu River tributary, to the final stop at "the end of the road" at Tarndale Gully (Fig. 1).

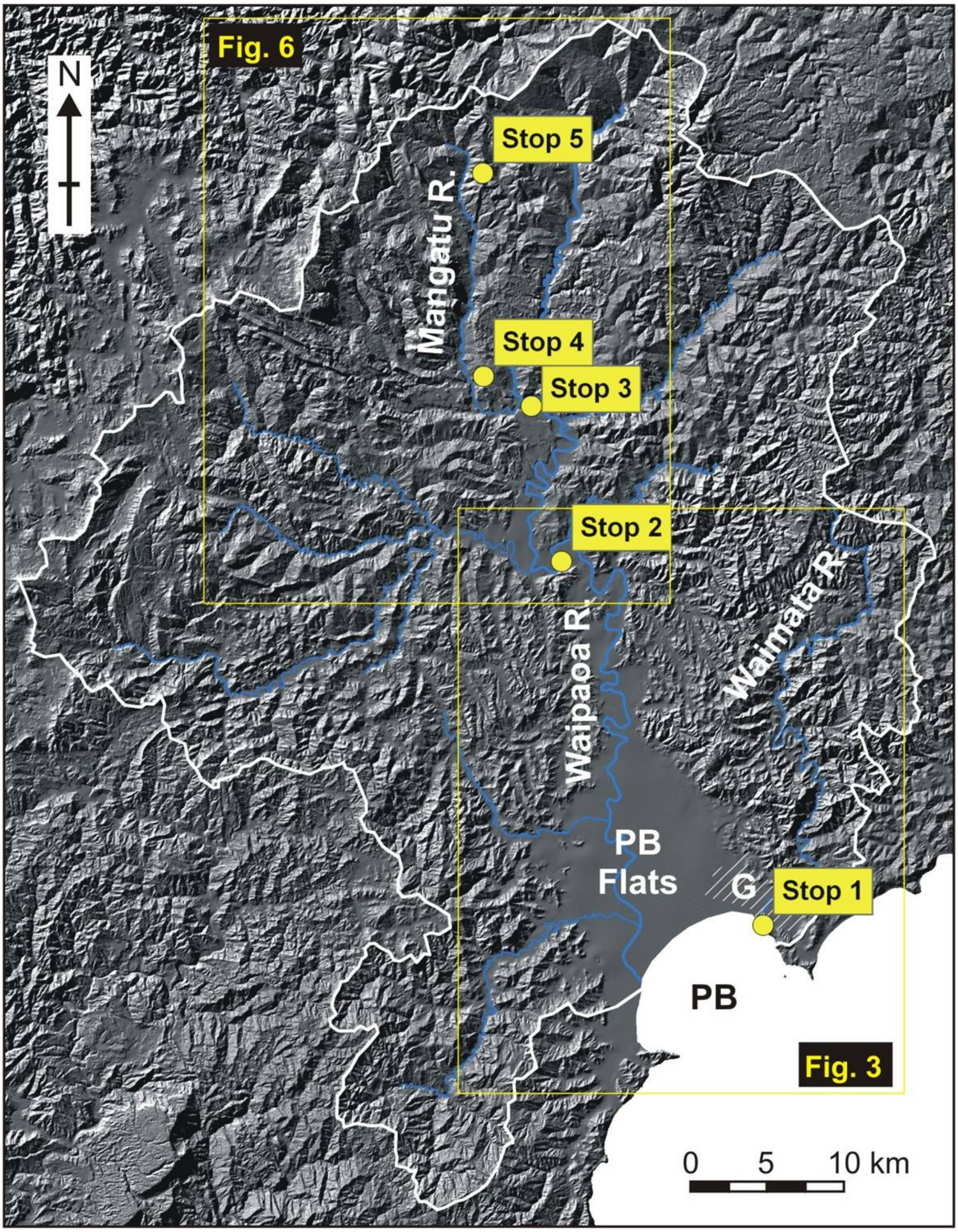

Fig. 1 Shaded Digital Terrain Model showing the fieldtrip stops. Blue lines are the major rivers/tributaries. White line is the Waipaoa-Waimata River catchment boundary. G = Gisborne, PB = Poverty Bay.

The Waipaoa SS is situated within the active forearc of the northern Hikurangi Subduction Margin, where oceanic crust of the Pacific Plate is being obliquely subducted beneath continental crust of the Australian Plate (Fig. 2). The northern sector of the Hikurangi Sunduction Margin is characterised by tectonic erosion, which has resulted in a narrow, steep forearc, the toe of which is indented as a result of numerous seamount impacts (Lewis and Pettinga, 1993; Collot et al., 1996). Tectonic erosion and sediment subduction has also resulted in significant (≤ 20 km thick) sediment underplating beneath the Raukumara Range (Reyners et al., 1999; Eberhart-Phillips and Chadwick, 2002).

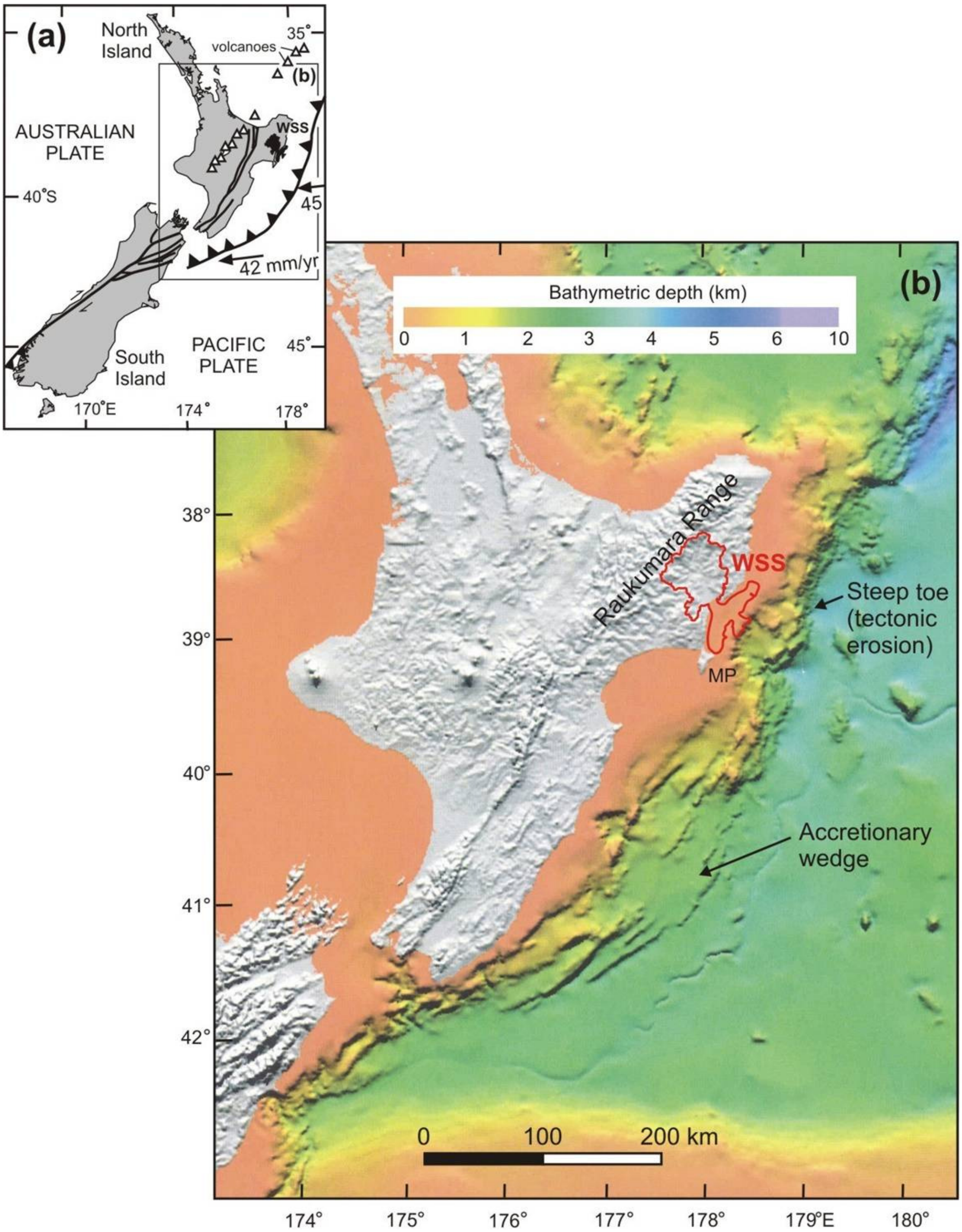

Fig. 2 **(a)** Plate tectonic setting of the Waipaoa SS (WSS). Plate convergence rates are from Wallace et al. (2007). **(b)** Shaded bathymetry map (CANZ Group, 1996) showing the variability along-strike of the Hikurangi Subduction Margin. MP = Mahia Peninsula.

The offshore portion of the Waipaoa SS, the Poverty Bay shelf and slope, straddles the upper part of the accretionary prism and the forearc basin. The Neogene accretionary prism is characterised by high slip rate (several mm/yr) thrust faults (P. Barnes and J. Mountjoy unpubl. data), which have uplifted Miocene mudstone-cored anticlines (Ariel and Lachlan). These anticlines bound subsiding sedimentary basins which have been rapidly infilled due to the high input of terrigenous sediment (Foster and Carter, 1997; Orpin et al., 2006; Gerber et al., in review).

The terrestrial portion of the Waipaoa SS, the Waipaoa River catchment, is situated upon mudstone-dominated Miocene-Pliocene forearc basin sediments (Mazengarb and Speden, 2000). Uplift is generally occurring on a regional scale, inferred to be the result of deep-seated subduction processes such as sediment underplating (Walcott, 1987; Litchfield et al., 2007). The relatively few active faults in the Waipaoa River catchment are low slip rate (<1 mm/yr) normal faults (K. Berryman and M. Marden, unpubl. data), which are interpreted to be secondary faults formed in response to the rapid uplift (i.e., gravitational collapse) of the Raukumara Range (e.g., Thornley, 1996).

The Waipaoa SS is characterised by a temperate maritime climate, with a relatively large number of sunshine hours and low wind speeds, but is punctuated by tropical cyclones (Hessell, 1980). The predominantly westerly winds result in orographic mean annual rainfall varying from 1000 mm at the coast to 3000 mm in the headwaters, with the greatest rainfall occurring in winter (Hessell, 1980). The paleoclimate record is patchy, but pollen records indicate that the early Holocene was wetter and warmer than present (Mildenhall and Brown, 1987), and that fully developed forest existed in the Gisborne region during all but coldest parts of the glacial periods, when it is likely that grass or shrublands characterised (non-glaciated) hillslopes undergoing freeze-thaw processes (McGlone et al., 1984). Deforestation commenced ~700 years ago during early Maori settlement (Wilmshurst et al., 1999), but increased dramatically following European settlement in the 1830s (MacKay, 1982) with major clearance in the headwaters occurring in the period 1880–1920 (Henderson and Ongley, 1920; Mackay, 1982). The rapid recent deforestation has had dramatic impacts on the landscape and sediment delivery of the Waipaoa and nearby river catchments (e.g., Marden et al., 2005; Gomez and Trustrum, 2005; Orpin et al., 2006; Gomez et al., 2007).

Stop 1: Kaiti Hill Lookout

The main Kaiti Hill Lookout offers views across Poverty Bay and the lowermost Waipaoa River catchment (Fig. 3). On a clear day, Mahia Peninsula can be viewed off in the distance to the south, which is being actively uplifted by the Lachlan Fault. In the foreground is Poverty Bay, bound on the southwest side by Young Nicks Head, named by Captain James Cook after 12 year old Nicholas Young, who in 1769 was the first European to sight New Zealand since Abel Tasmans voyage in 1642. The Waipaoa River mouth is difficult to see, but is currently situated at the far (west) end of the actively prograding sandy beach (Smith, 1988). Gisborne city is situated at the eastern corner of the triangular shaped coastal plain, the Poverty Bay Flats, which are actively tilting westward and trapping substantial volumes of Waipaoa River sediment (Brown, 1995; Wolinsky et al., in review) (Q1a on Fig. 3). The surrounding hills are predominantly Miocene and Pliocene mudstones and sandstones of the Tolaga (Ml) and Mangaheia (Pm) Groups (Mazengarb and Speden, 2000). These are locally overlain by predominantly lacustrine sediments of the early-middle Pleistocene Mangatuna Formation (Neef et al., 1996; Kennedy et al., 2008) (eQa) and in the middle and upper catchment, late Pleistocene fluvial terraces (Berryman et al., 2000; Marden et al., 2008) (Q2a, Q3a, Q4a, Q5a).

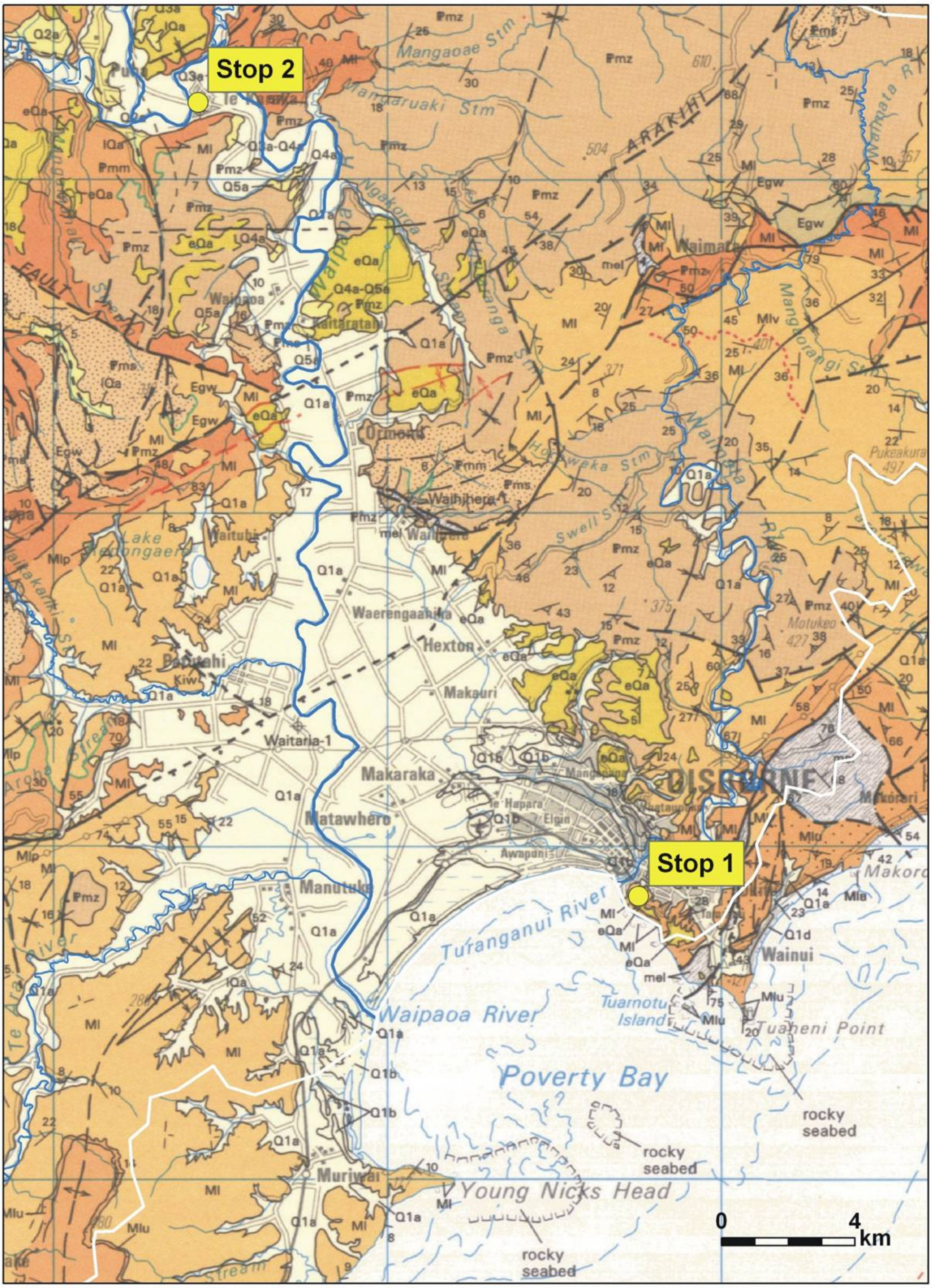

Fig. 3 Geological map (Mazengarb and Speden, 2000) showing the locations of stops 1 and 2. Unit coding is based primarily on age; M = Miocene, P = Pliocene, Q = Quaternary. Specific important units are referred to in the text for stop 1. Thicker blue line is the Waipaoa River mainstem. White line is the Waipaoa-Waimata River catchment boundary.

Poverty Bay

The seaward end of the land-sea interface is a shallow, bowl-shaped bay approximately 8 km wide, headland to headland, and 5 km in length from the shoreline to the open ocean. Maximum depths reach 25 m at the mouth of the bay, sloping very gently towards the shoreline before reaching the steep, nearshore shoreface (Fig. 4). Poverty Bay is largely wave-dominated with weak astronomical tides (2-4 cm/s) and wave heights within the bay averaging 1.2m and storm waves of 4-6 m (Bever, personal communication).

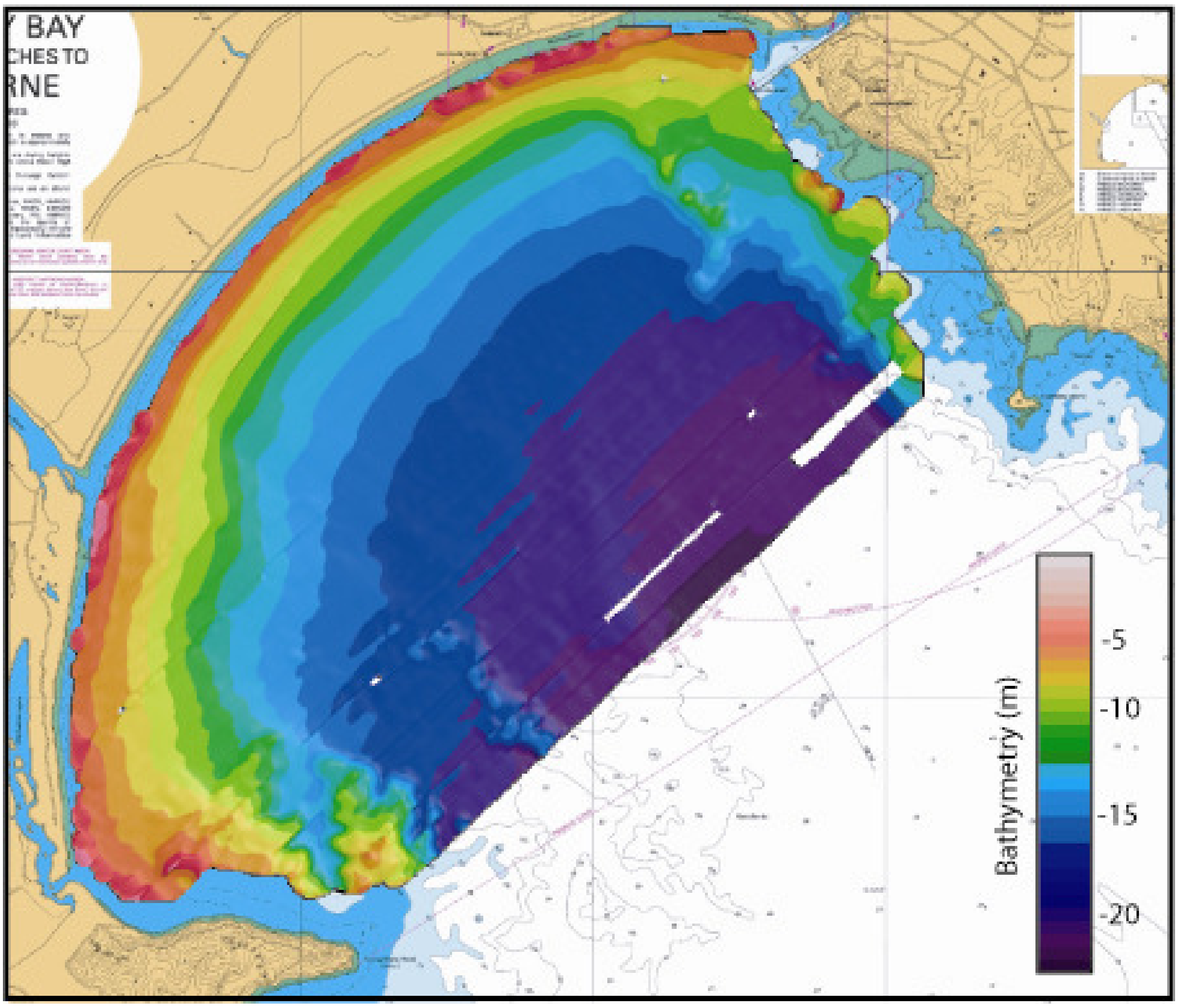

Fig. 4. Bathymetric chart of Poverty Bay collected with an interferometric swath system in August 2005.

The high wave energy combined with the isolation of the littoral zone by the enclosing headlands creates very efficient segregation processes leading to capture of sand-sized sediment within the prograding shoreline. The fine-grained sediment, conversely, appears to be efficiently bypassed over the long-term across the bay to the open ocean. We suspect the seaward region of Poverty Bay is virtually full with respect to accommodation space when one considers the depth of strong, reworking wave energy and the near exposure of rock (Fig. 5). Chirp seismic profiles collected across Poverty Bay show a sill of rock exposed or just below the surface across most of Poverty Bay.

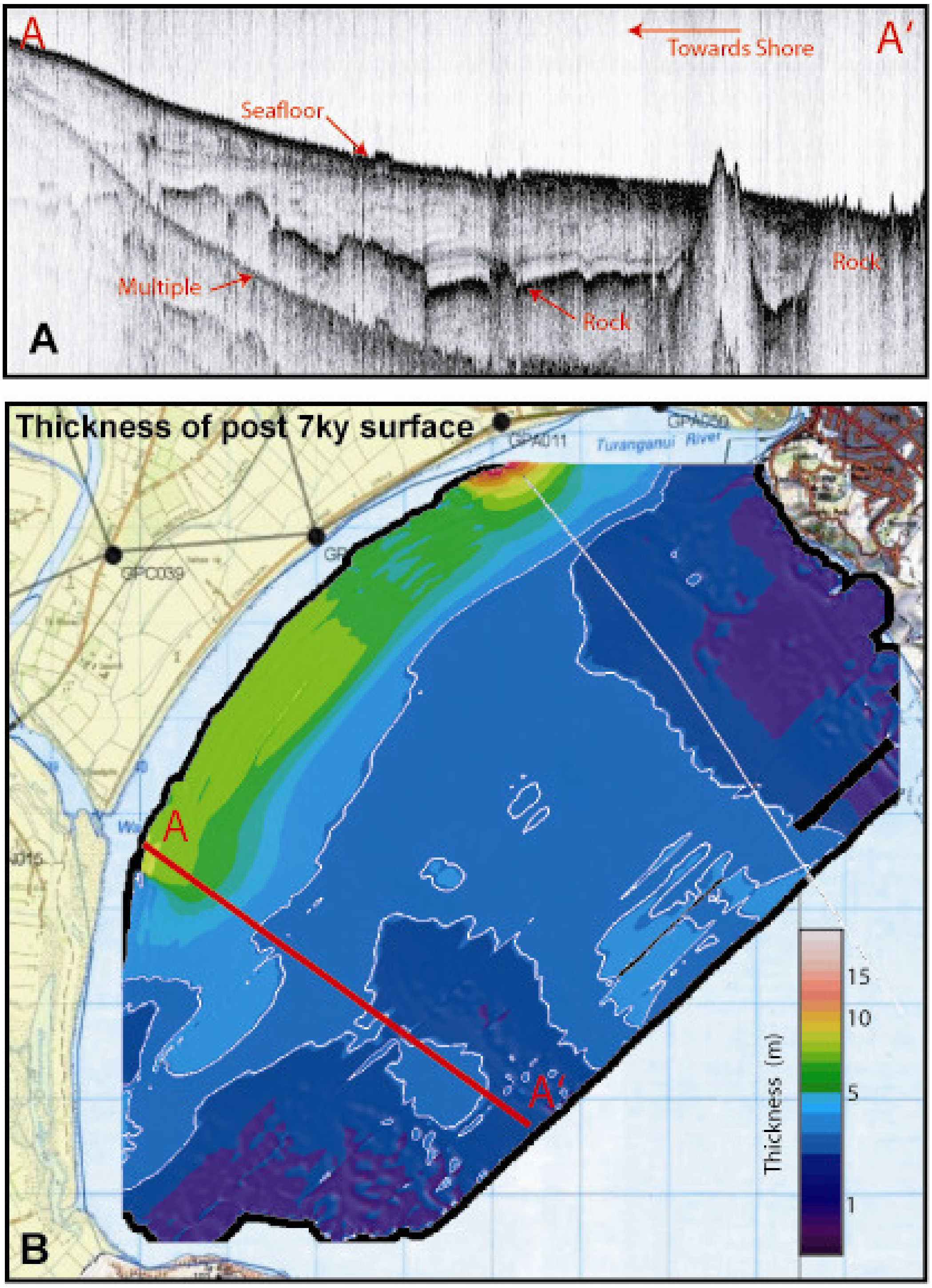

Fig. 5. Chirp seismic profiles across Poverty Bay, exemplified in A-A’, show rock near the surface of the seafloor and minimal sediment preserved above.

Chirp seismic profiles groundtruthed by Poverty Bay vibracores and adjacent coastal plain boreholes (Litchfield, personal communication) reveal a transgressive surface (~7 ka) extending from the coastal plain out across Poverty Bay (Fig. 6). Within Poverty Bay, this transgressive surface is comprised of a thin (~10 cm) clay lens overlain by coarse shelly hash. A thin (< ~3 m) veneer of sandy sediment overlies this surface, and is thickest along the shoreline and within the nearshore (~10-15 m, Fig. 5b). Vibracores indicate that sediments preserved above this surface are dominantly sandy, suggesting significant re-working and bypassing of fine-grained material.

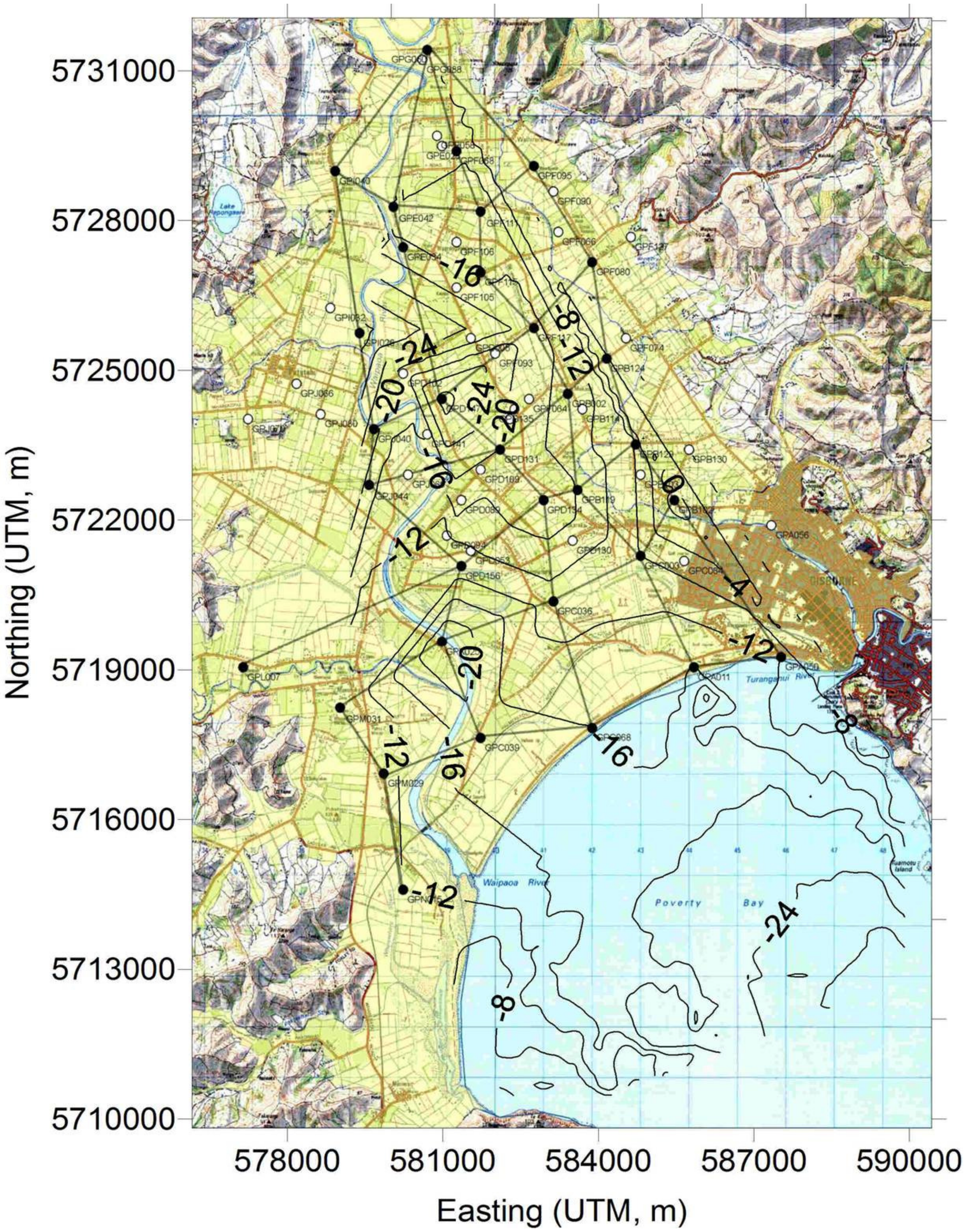

Fig. 6. Contour (m, msl) map of ~7 ka transgressive surface compiled from chirp seismic profiles, vibracores and coastal plain boreholes.

Poverty Bay Flats

On the Poverty Bay Flats the Waipaoa River has a meandering channel with active migration and numerous cutoffs documented in the historical period. Over the late Holocene the lower Waipaoa River migrated westward across Poverty Bay Flats, possibly driven by tectonic steering, with the Waipaoa River mouth migrating from near Gisborne harbor, ~1480 BC, to its present position, by ~1888 AD (Pullar and Penhale, 1970). Previous positions of the river mouth are marked by several lagoons which interrupt the smooth sandy shoreline. The modern floodplain is aggrading rapidly due to the pulse of erosion triggered by deforestation (Gomez et al., 1998). The weakly concave form of the river profile suggests the coastal plain has a quasi-equilibrium form which balances fluvial sediment transport with long-term shoreline progradation and tectonic subsidence (Sinha and Parker, 1996). Post-LGM transgression completely inundated Poverty Flats, but since sea level stabilized (~7 ka) the Poverty Bay shoreline has prograded over 12 km to reach its present position. Paleo-shorelines extrapolated from borehole logs and beach ridges (Fig. 7) show a systematic decelleration in progradation over this period (Brown, 1995), a consequence of increasing sediment storage by subsidence and aggradation over the growing alluvial plain (Wolinsky et al., in review).

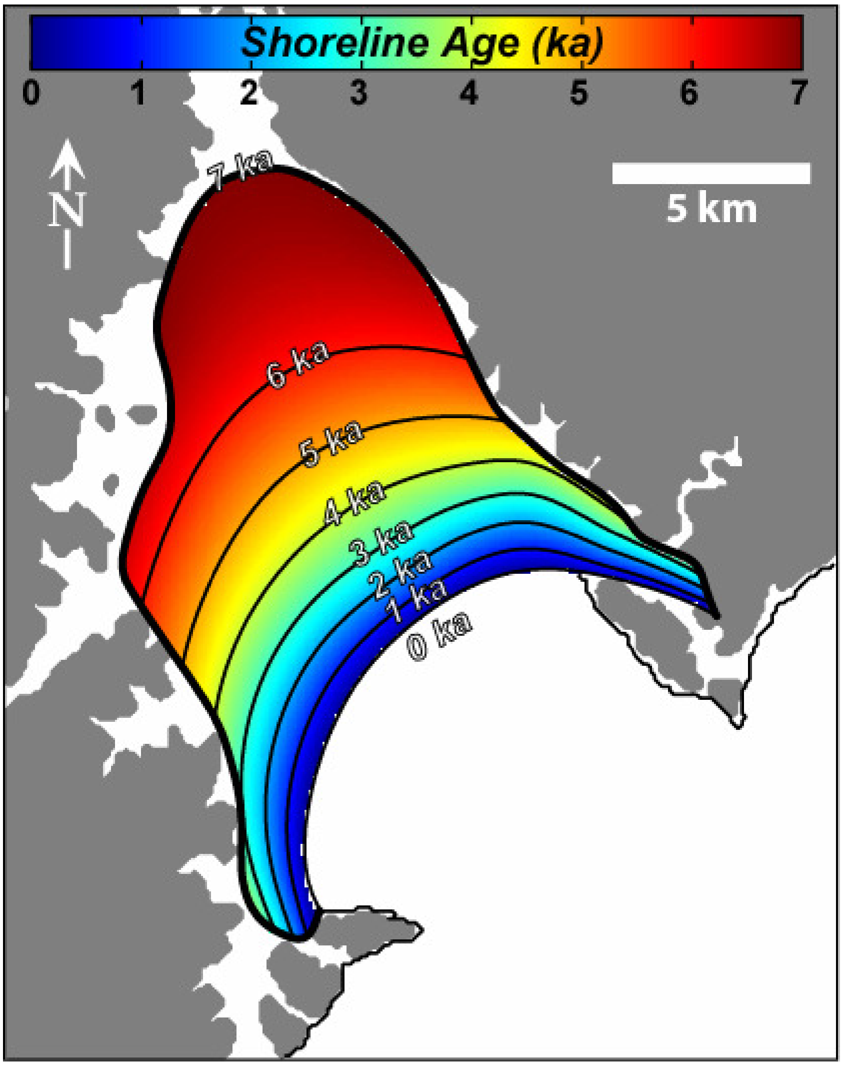

Fig. 7. Holocene progradation of the Poverty Bay shoreline, reconstructed from beach ridges, borehole stratigraphy, and carbon-dated shells (Wolinsky et al., in review).

Stop 2: Te Karaka (comfort stop)

Te Karaka is situated approximately in the middle of the Waipaoa River catchment (Fig. 1), so is a convenient place for a comfort (toilet) stop.

Te Karaka is situated on relatively young (post-18 ka) fluvial terraces of the Waipaoa River. Remnants of older terraces are preserved at slightly higher levels to the north and south (Berryman et al., 2000). Approximately 3 km further along the road we turn off state highway 2 and drive north upon a series of diverging and rapidly climbing fluvial terraces. These will be discussed at Stop 3 (Whatatutu).

Stop 3: Whatatutu School

Whatatutu is situated near the junction of the Waipaoa and Mangatu Rivers (Fig. 8), above which the Waipaoa River flows through a narrow gorge. A change in geology also occurs at this point, with the units to the west forming part of the East Coast Allochthon (hatch units in Fig. 8), which will be discussed at Stop 4 (Mangatu Rd).

Whatatutu Township and School are situated on a Waipaoa River fluvial terrace formed during the Last Glacial Maximum, locally referred to as Waipaoa 1, or W1 (Berryman et al., 2000; Marden et al., 2008). W1 is the lowermost of a series of fluvial terraces (W1-W4) inferred to have formed during cold periods in response to reduced vegetation and enhanced periglacial processes (Berryman et al., 2000), as has been described elsewhere in the North Island (Litchfield and Berryman, 2005; Clement and Fuller, 2007). Thus the W1-W4 fluvial terraces record periods of prolonged river aggradation, whereas the spacing between them represents periods of river downcutting or incision, which is in part driven by uplift (Berryman et al., 2000; Litchfield and Berryman, 2006).

The ages of fluvial terraces in the Waipaoa River catchment are relatively well constrained from their tephra and loess coverbed stratigraphy (Berryman et al., 2000; Eden et al., 2001). The oldest tephra overlying W1 fluvial gravels is the 17.7 cal. ka Rerewhakaaitu Tephra (Berryman et al., 2000; Eden et al., 2001; Marden et al., 2008; Berryman et al., in review), suggesting it was abandoned at approximately the end of the last glacial coldest period (~18 ka; Alloway et al., 2007). The age of the base of the W1 terrace deposits, which are up to 30 m thick, is poorly constrained, but the presence of the 27 cal. ka Kawakawa Tephra within gravels in a few localities (Marden et al., 2008; Berryman et al., in review) suggests it is likely to be approximately the start of the last glacial coldest period (~28 ka; Alloway et al., 2007).

The W1 terrace is widely preserved throughout the Waipaoa River catchment (Marden et al., 2008), and so provides an invaluable marker for landscape and sediment budget studies. Marden et al. (2008) used the W1 surface to calculate the volume of sediment eroded from river channels in the last 18 ka by reconstructing the former extent of the W1 terrace surface in a GIS and then subtracting the modern day channel surface (Fig. 9). The volumes calculated are ~9.5 km^3 for the Waipaoa River catchment and ~2.6 km^3 for the adjacent Waimata River catchment. Marden et al. (2008) also calculated the volume of post-18 ka sediment currently stored beneath the Poverty Bay Flats flood plain, from the thickness of sediment above a gravel aquifer interpreted to be the (now subsided and buried) correlative of the W1 terrace. The calculated volume is ~6.6 km^3, and thus the difference between these two estimates, ~5.5 km^3, is the volume of river channel sediments delivered to the ocean. This amounts to approximately one third of the ~18 km^3 of post-18 ka sediment stored on the shelf and slope basins (Orpin et al., 2006), and the remainder is interpreted to be sediment directly delivered to the ocean from hillslope erosion.

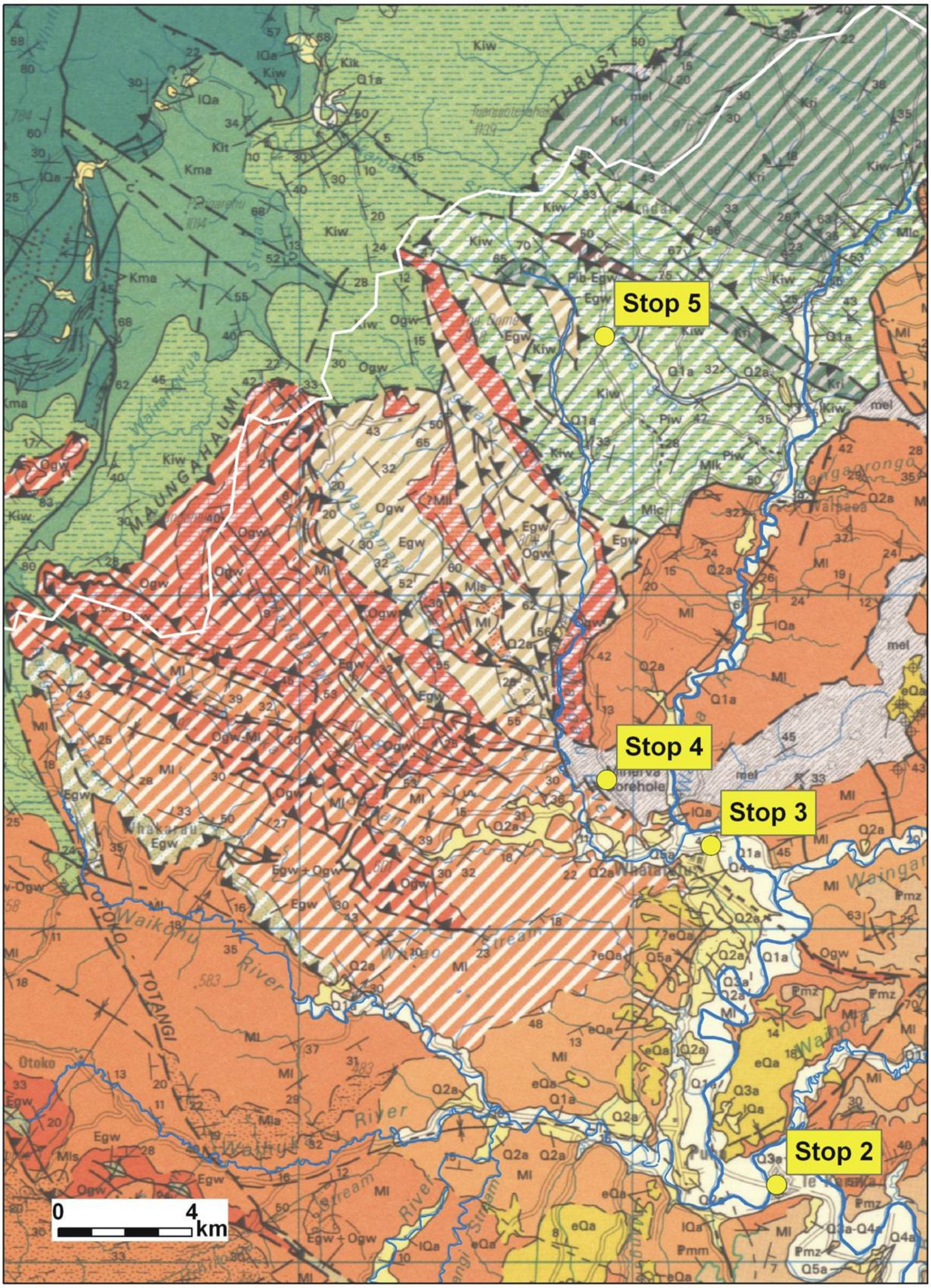

Fig. 8 Geological map (Mazengarb and Speden, 2000) showing the locations of stops 2 to 5. Unit coding is based primarily on age; K = Cretaceous, E = Eocene, O = Oligocene, M = Miocene, P = Pliocene, Q = Quaternary. Hatch pattern denotes the early Miocene East Coast Allochthon. Thicker blue line is the Waipaoa River mainstem, the Mangatu River tributary is to the west. White line is the Waipaoa-Waimata River catchment boundary.

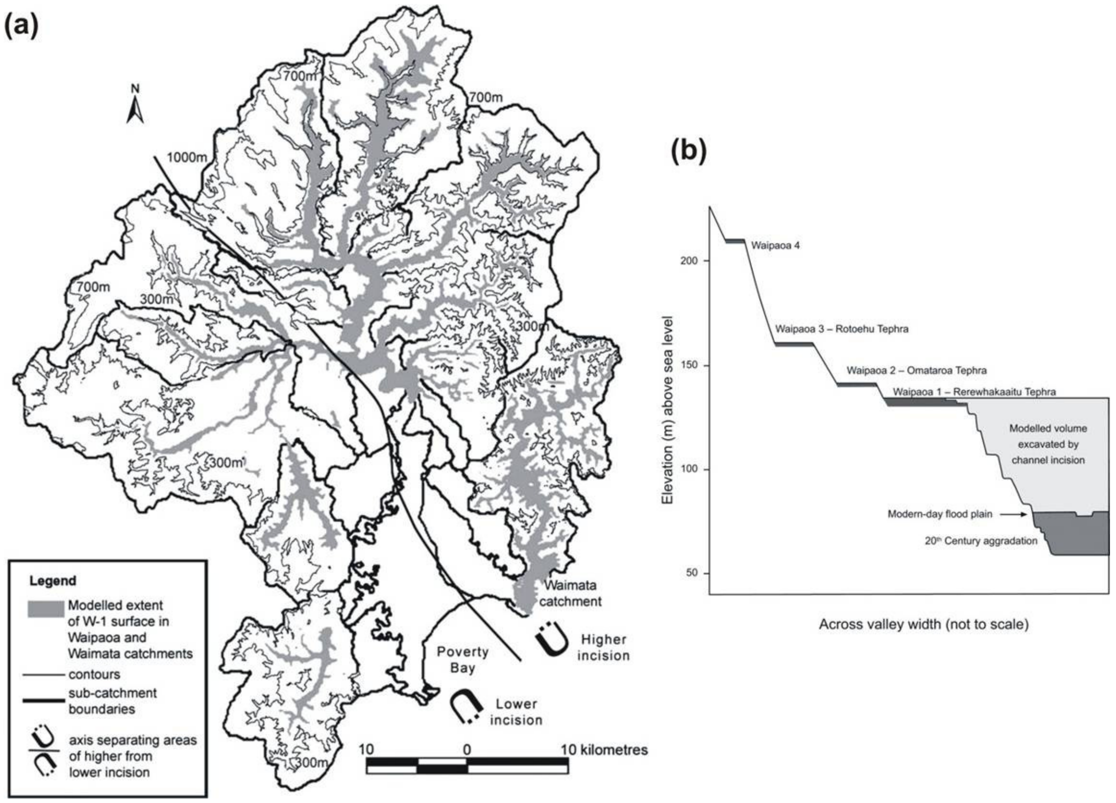

Fig. 9 **(a)** Map of the reconstructed (modelled) W1 surface within the Waipaoa and Waimata River catchments (Marden et al., 2008). Note the higher incision in the NE. **(b)** Schematic cross-section showing the channel area eroded below the W1 terrace (Marden et al. 2008).

Details of the rates and volumes of erosion of the river channels below W1 is the subject of ongoing study. Crosby (2006) proposed that the mechanism of channel erosion is primarily knickpoint (discrete steep steps in the riverbed producing a convexity in the river profile, Fig. 10) retreat. Crosby and Whipple (2006) showed that the majority of major knickpoints are currently situated in the uppermost reaches of the Waipaoa River catchment, at drainage areas between 1×10^5 m^2 and 1×10^6 m^2. Although their numerical modelling was unable to distinguish between knickpoint initiation at the river mouth or at major tributary junctions, they were extremely effective in demonstrating that the valley floor above a knickpoint preserves "relict topography" which projects downstream to the W1 terrace (Fig. 10). That is, the area above the knickpoints are hanging valleys which are yet to undergo major post-glacial incision and therefore the surrounding hillslopes have not yet felt the effects of relative baselevel drop (this will be discussed further at Stop 4). Berryman et al. (in review) have studied a flight of post-18 ka (i.e., below W1) degradation (strath) terraces in the Waihuka River tributary, and showed that a period of rapid downcutting in the early Holocene (~10-8 ka), which corresponds to a steepening in terrace gradient, may record the passing of a knickpoint (or a knickzone).

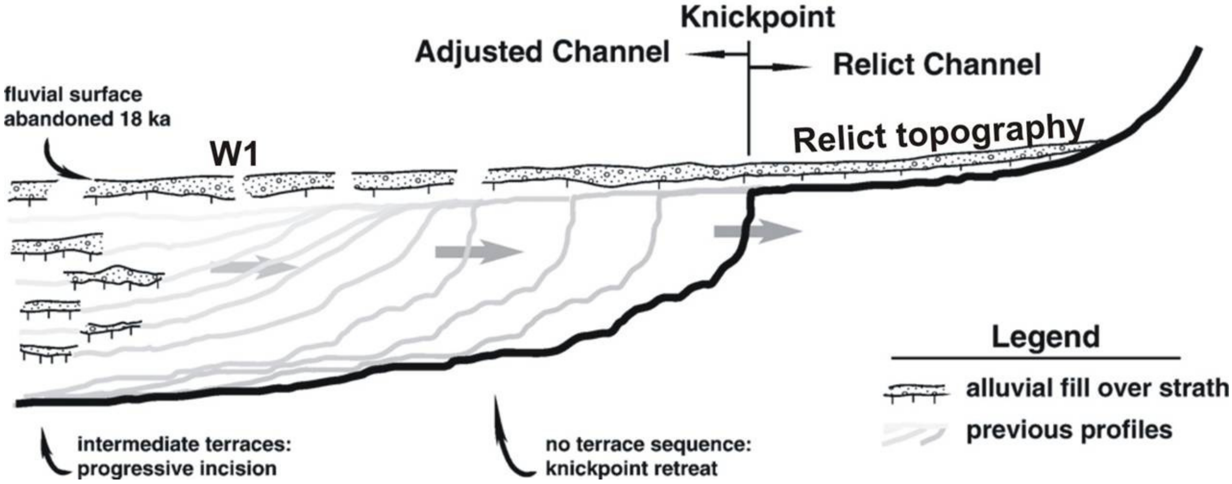

Fig. 10 Schematic illustration of a typical knickpoint on a long profile of the Waipaoa River (modified from Crosby and Whipple, 2006). The flight of terraces in the Waihuka River tributary studied by Berryman et al. (in review) indicates that incision may occur by a combination of progressive incision (l.h.s.) driven by tectonic uplift and knickpoint retreat (centre).

Stop 4: Mangatu Road

This roadside viewpoint looks out across the west side of the Mangatu River catchment, which is characterised by fluvial terraces and large, deep-seated bedrock landslides. This area is underlain by a series of nappes of Oligocene, Eocene, and early Miocene mudstones and sandstones of the East Coast Allochthon (hatch units in Fig. 8) (Mazengarb and Speden, 2000). The East Coast Allochthon was emplaced in several stages during the early Miocene, and is considered to have been driven by delamination and obduction of a flake of the uppermost oceanic plate during the initiation of subduction (Rait et al., 1991; Field, Uruski, and others, 1997). The northwest strike of the nappes exerts a strong structural grain on the topography, along which a series of tributaries are guided.

As noted at the previous stop, the aggradation river terraces, also visible at this viewpoint, provide an invaluable marker for landscape and sediment budget studies. Sediment budget work by Marden et al. (2008) suggested that as yet unquantified hillslope erosion process have been significant contributors to the total postglacial sediment yield. Deep-seated landslides are pervasive in the landscape of the Waipaoa River catchment and represent one such hillslope erosion process. Deep-seated mass movement in the catchment includes rotational, translational and extensive earthflow landslides. Work currently under way in two geologically distinct tributary areas of the Waipaoa River catchment (Fig. 11) has the goal of developing methodologies for quantifying the postglacial sediment flux from deep-seated landslides in the catchment.

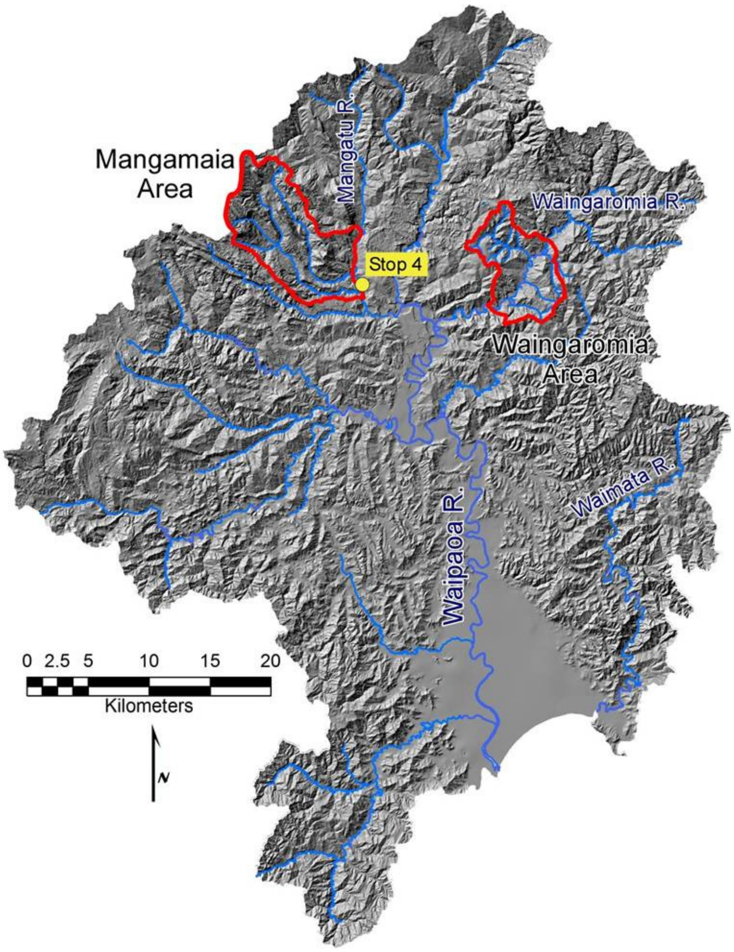

Fig. 11 The Waipaoa River catchment with some major rivers labelled. Red outlined areas are the two tributary areas where the role of deep-seated landslides in the Waipaoa SS is currently being investigated. The Waingaromia area on the east side of the catchment is underlain by mostly Miocene mudstones of the Tolaga group and some Miocene melange (Mazengarb and Speden, 2000). The Mangamaia area on the west side of the catchment is underlain by Oligocene to Early Miocene allochthonous mudstones and limestones of the Tolaga and Mangatu groups (Mazengarb and Speden, 2000).

Recognized deep-seated landslides that are directly connected with the fluvial system comprise 20-30 percent of the surface area of the two sections under study (Fig.12). One potential mechanism for such widespread instability is removal of an effective hillslope toe buttress system by the widespread post-18 ka channel incision. Preliminary work suggests that many of these large landslides are downstream of knickpoints in their respective tributaries and, based on tephras cover on landslide debris, have been active in the Holocene.

To calculate the sediment production from these landslides and to differentiate this sediment source from channel incision, two topographic reconstructions need to be accomplished: (i) the early post glacial landscape around the landslides needs to be reconstructed, as in Marden et al. (2008), and (ii) pre-landslide topography needs to be interpolated from geomorphic features such as head and side scarps. The limiting factors for such reconstructions are the preservation of landslide scarps and the resolution of the modern topographic data set (subtracting surface).

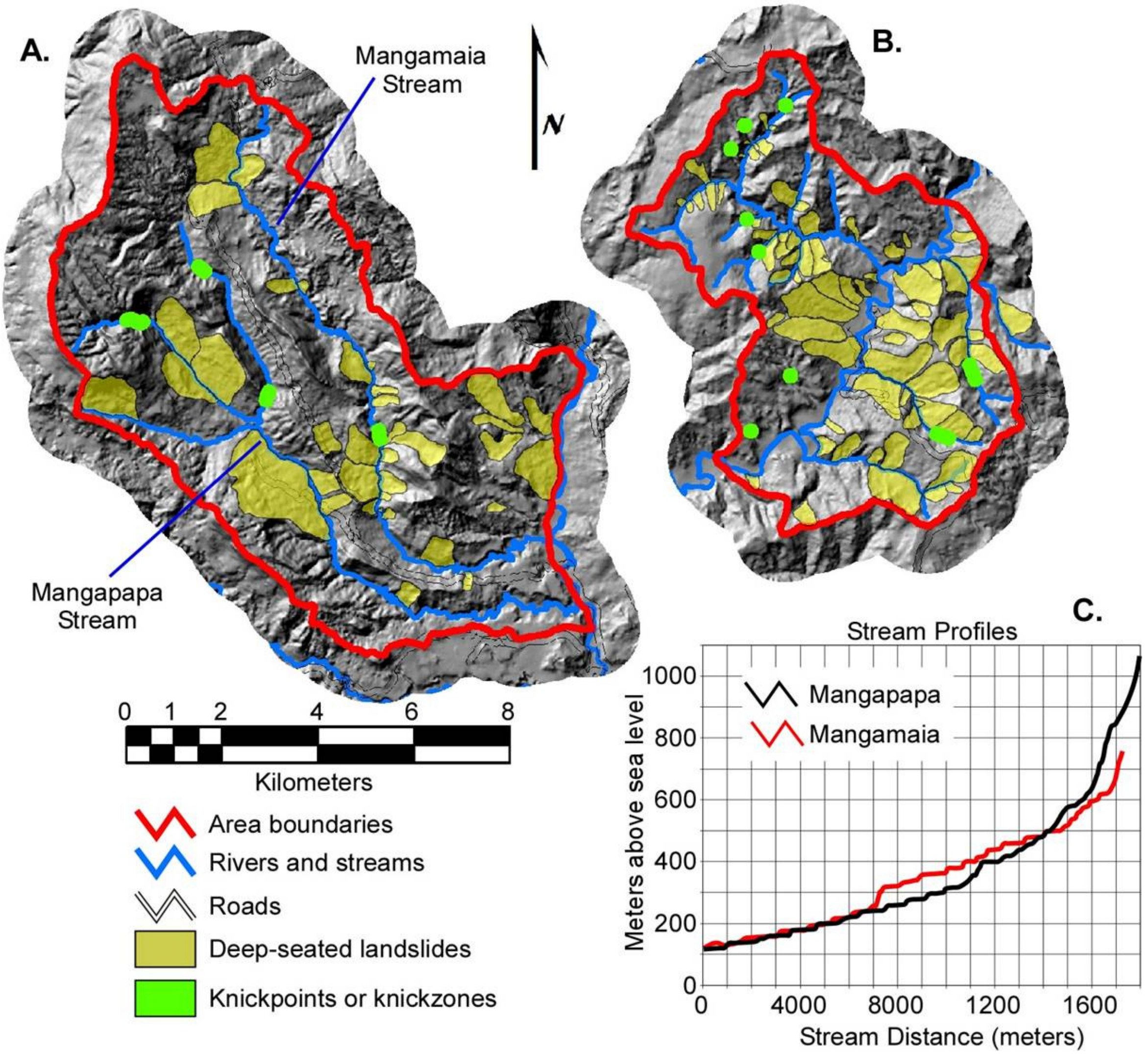

Fig. 12 Two tributary areas where the role of deep-seated landslides in the Waipaoa SS is currently being investigated. Deep-seated landslides and knickpoints mapped from a combination of air photos and field work. Also shown are select stream profiles. **A.** The Mangamaia area drains about 7450 ha; it is characterized by high relief in the uplands and is underlain by allochthonous mud and limestones. Major knickpoints occur approximately at the midpoint of the two main tributary streams in this area. **B.** The Waingaromia area drains about 4400 ha, is more moderate in relief and underlain by mudstones and, in places, low-strength melange. In this area, knickpoints are in the uppermost reaches of even the smallest tributaries. **C.** Stream profiles of the two main tributary streams in the Mangamaia area showing some of the knickpoints in profile, shown on map A.

Stop 5: Tarndale Gully

Background

Anecdotal evidence suggests that mass movement and gullying was initiated in the winter of 1915 in response to the removal of the forest cover with subsequent loss of root strength lowering the threshold for erosion. The slopes surrounding Tarndale Gully were replanted in 1963. While reforestation, through a combination of reduced runoff and a shortened period of soil moisture surplus ameliorated erosion on the lower slopes surrounding Tarndale Gully, the large area of bare gully-head was far too steep and active to be mitigated by afforestation and it remains active today.

The watershed surrounding Tarndale Gully at its highest elevation is 580 m. At its upstream end four tributary fans each draining a different part of the ~13 ha, amphitheatre-shaped, gully headwall converge to form a central channel/fan.

The channel/fan extends from the base of the headwall (elevation 420 m) to its junction with Te Weraroa Stream (elevation 320 m), a distance of ~1 km.

Past trends in sediment storage and transfer

Digital elevation models (DEMs) were derived from sequential aerial photographs (1939, 1958, 1992 & 2004) to quantify the amount of sediment stored in the gully-fan-channel complex. Decadal changes in fan storage derived from DEMs were compared with cross-section surveys of the channel/fan for the period 1950-83. Between 1950 and1960, aggradation increased by ~27 000 m^3/yr as Tarndale Gully expanded, but declined to ~12 000 m^3/yr during the 1960-83 period when sediment supply slowed largely in response to reforestation. Since 1983 biannual cross-section measurements of the active and then incised channel/fan show for the period 1983-2006 that changes in storage have a seasonal pattern. Channels infill during winter (May-October) when debris flows, generated on the steep flanks of the headwall, transports sediment en masse to the channel/fan (Fig. 13b). In contrast, channel incision occurs during periods of sediment starvation which coincide with low flow in the summer (November-April) (Fig. 13a). Since 1983, seasonal aggradation/degradation cycles in channel/fan storage indicate a slowing in net aggradation to ~1700 m^3/yr and, more recently (1995-2003), a net storage loss of ~5000 m^3/yr. This reflects slowing sediment production from the headscarp and a consequent net export of sediment from the channel/fan to Te Weraroa Stream. Since 2005, following significant rainfall and flooding, mass movement activity in the gully headwall has increased resulting in net aggradation of ~35 000 m^3/yr elevating the channel/fan to a level not reached for more than 2 decades.

Current research on trends in sediment storage and transfer

Connectivity between the Tarndale Gully and the upper Waipaoa catchment which it feeds via a steep alluvial fan (Fig. 13c) has been assessed by joint research between Massey University and Landcare Research utilising DEMs of the Tarndale Fan derived from ten RTK-dGPS surveys between 2004 and 2008. Repeat DEM analysis permits an assessment of the transfer of sediment supplied from the gully through the fan to the Te Weraroa Stream. This indicates the operation of two critical junction switches at the (i) gully-fan and (ii) fan-stream nexus which contribute to considerable complexity in patterns of erosion and deposition on the fan as it responds to sediment supply or starvation from the Tarndale Gully and evacuation to the Te Weraroa Stream. The fan does not respond as a coherent unit. Each of five fan feeder tributaries behaves independently according to quantities of sediment supplied from discrete source areas within the gully complex. This enables detection of activity within specific components (activity zones) of the Tarndale Gully. Mass movements, triggered by single rainstorm events, wet periods and intrinsic mechanisms, frequently deliver sediment to the upper fan. Extreme rainfall events are particularly effective at delivering large quantities of sediment. Enhanced sediment delivery promotes rapid aggradation (up to ~30 000 m^3) of the upper fan as a whole, with each feeder tributary responding to its discrete source zone. Infilling may subsequently propagate down-fan, particularly when the upper fan equally rapidly incises these deposits, transferring sediment as far as the Te Weraroa Stream. Floods generated in the Te Weraroa Stream by the same extreme rainstorm may trim the lower fan prompting up-fan incision in response to changed local base level. This process may evacuate up to ~10 000 m^3 of coarse sediment to the stream. This see-saw behaviour in response to sediment supply and evacuation is superimposed on the recent overall aggradational trend in which the Tarndale Fan is buffering the trunk channel system from the full quantity of sediment supplied by the Tarndale Gully.

For more information see poster entitled: 'Slope-channel coupling at a critical nexus in the Waipaoa sediment cascade: Tarndale Gully & Fan, Mangatu Forest, New Zealand'. Fuller, I. C. (Massey University) & Marden, M. (Landcare Research NZ Ltd).

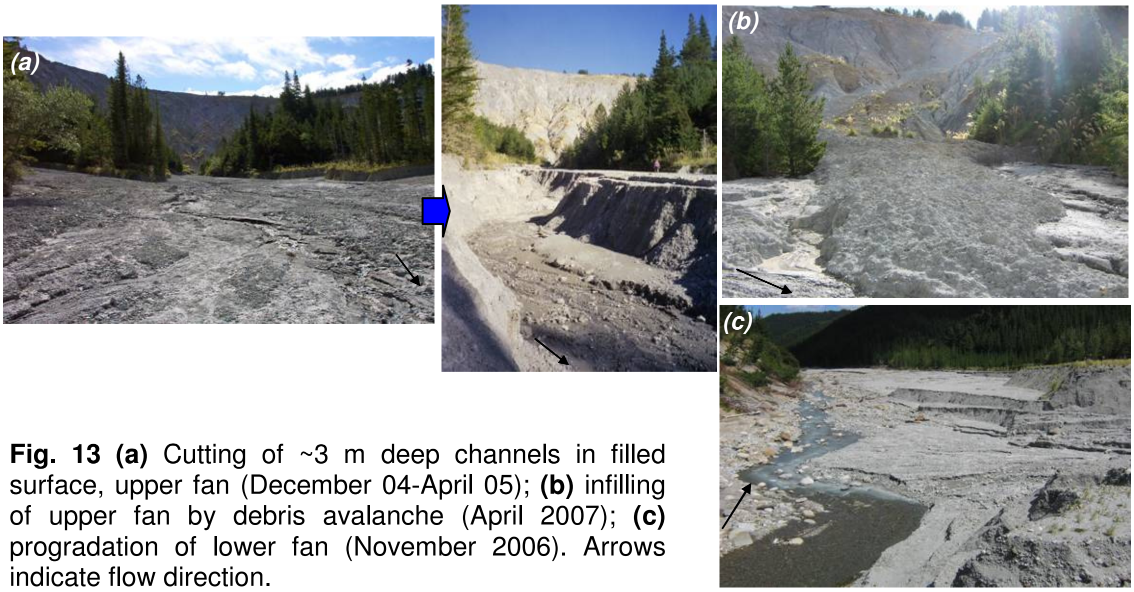

Fig. 13 **(a)** Cutting of ~3 m deep channels in filled surface, upper fan (December 04-April 05); **(b)** infilling of upper fan by debris avalanche (April 2007); **(c)** progradation of lower fan (November 2006). Arrows indicate flow direction.

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www.gns.cri.nz

Principal Location

1 Fairway Drive
Avalon
PO Box 30368
Lower Hutt
New Zealand
T +64-4-570 1444
F +64-4-570 4600

Other Locations

Dunedin Research Centre
764 Cumberland Street
Private Bag 1930
Dunedin
New Zealand
T +64-3-477 4050
F +64-3-477 5232

Wairakei Research Centre
114 Karetoto Road
Wairakei
Private Bag 2000, Taupo
New Zealand
T +64-7-374 8211
F +64-7-374 8199

National Isotope Centre
30 Gracefield Road
PO Box 31312
Lower Hutt
New Zealand
T +64-4-570 1444
F +64-4-570 4657