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1 **Contrasting mixed siliciclastic-carbonate shelf-derived gravity-**  
2 **driven systems in compressional intra-slope basins**  
3 **(southern Hikurangi margin, New Zealand)**

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## 22 1. ABSTRACT

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23 Along active margins, the combination of predominant tectonic activity and shallow-marine mixed  
24 siliciclastic-carbonate source systems developing upon and around actively growing structures  
25 challenges traditional source-to-sink models. This study aims to investigate the implications of mixed  
26 siliciclastic-carbonate shelfal domains located in contrasting geotectonic settings (thrust forelimb and  
27 backlimb) for the development of the concomitant gravity-driven systems beyond the shelf edges. Here,  
28 we document the vertical and lateral stratigraphic variabilities of the shelf-derived turbidites and mass-  
29 transport deposits (MTDs) at outcrop-scale through the integrated interpretation of photogrammetry, field  
30 and taphonomic data from the emerged southern portion of the Hikurangi subduction margin. Results  
31 highlight the role and importance of varying structural setting of the sediment source, whereby the  
32 different morphologies of the source regions (continent-attached forelimb, continent-detached backlimb)  
33 control the development of highly varied shelf-derived gravity-driven depositional systems that interact  
34 with the structures across the same confined intra-slope basin. The deposits are tens to a few hundred  
35 of meters in thickness and have a lateral extent of several kilometers. The depositional systems are  
36 characterized by durations of 1 to 2 Ma and were primarily controlled by the geometries and tectonic  
37 motion of the underlying structures at the shelf edges. Shelf-derived mass-wasting systems occurred on  
38 both sides of the actively growing thrust structures and were sourced from both shelfal domains that were  
39 attached or detached from the continental domain. When sourced from the backlimbs however, the  
40 subsequent MTDs exhibit more complex internal architectures, ultimately recording the dynamic changes  
41 in slope gradient, and can therefore be used as proxies for unraveling the tectonic activity of an individual  
42 structure. Our study provides new insights to better predict mixed siliciclastic-carbonate depositional  
43 settings along active margins, sourced from thrust forelimb and backlimb. These results may be important  
44 for deep-marine exploration and tectonostratigraphic reconstruction of fold-and-thrust belts.

45 **Keywords:** active margin, confined slope channel system, mass-wasting system, mixed siliciclastic-  
46 carbonate systems, confined basins, trench-slope basins, thrust forelimb, thrust backlimb

## 47 2. INTRODUCTION

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48 Understanding sediment transfer and distribution beyond the shelf edge and in the outboard part of  
49 continental margins has been at the heart of numerous studies in the past 60 years (e.g., Bouma 1962;  
50 Posamentier and Vail 1988; Richards et al. 1998; Posamentier and Kolla 2003; Sømme et al. 2009;

51 Paumard et al. 2020). Commonly, the pathways taken by sediments across the shelf margins are believed  
52 to some extent control different types of sediment gravity flows and related depositional systems (Prather  
53 2003; Gong et al. 2016; Paumard et al. 2020). Submarine canyons and slope channels are more likely  
54 to funnel turbidity currents (Peakall and Sumner 2015), transferring continental and or shelf-derived  
55 material downslope (Kuenen 1964; Mulder et al. 2017) and normally ending their course as weakly  
56 confined to unconfined lobe systems (e.g., Brunt et al. 2013), whereas unconfined shelf edge failures  
57 typically generate shelf-derived debris flows that are either captured on the slope or deposited on the  
58 basin floor (e.g., Moscardelli and Wood 2008).

59 Along active margins, the combination of predominant tectonic activity and mixed siliciclastic-carbonate  
60 source systems developing upon and around actively growing structures challenges traditional source-  
61 to-sink models (Chiarella et al. 2012; Chiarella et al. 2019). Resulting deep marine sedimentation styles  
62 are highly variable, intricate and not readily predictable (McArthur et al. 2019). Whilst the underlying  
63 structural styles and motions fundamentally impact the morphology of continental shelves and their  
64 margins, they also induce dynamic changes in slope gradient, confinement and connection to sediment  
65 sources (Sømme et al. 2009). Shallow marine, mixed shelfal systems are common in active tectonic  
66 settings (Puigdefàbregas et al. 1986; Chiarella et al. 2012). Typified by repeated terrigenous sediment  
67 input (Chiarella et al. 2017), the shallow marine, mixed shelfal systems commonly result in intricate shelf  
68 sedimentation and sediment routing systems (Chiarella et al. 2019; Moscardelli et al. 2019; Cumberpatch  
69 et al. 2021). All these parameters control the dynamics of the flows from the shelf to the basin floor and  
70 can be expected to have a significant impact on the development of the coeval deep-marine systems,  
71 influencing (1) the types of sediment gravity flow, (2) the patterns of sediment dispersal, (3) the  
72 morphologies (size, extent) of the systems and (4) the related sedimentary facies (e.g., Fisher 1983;  
73 Pirmez et al. 2000; Mulder and Alexander 2001; Sprague et al. 2005; Sømme et al. 2009; Mulder et al.  
74 2012; Moscardelli et al. 2019; Paumard et al. 2020).

75 Although previous studies looked at the depositional frameworks of shallow-marine mixed siliciclastic-  
76 carbonate systems along tectonically active margins (e.g., Caron et al. 2004; Čosović et al. 2018;  
77 Chiarella et al. 2019), the stratigraphic implications for the subsequent shelf-derived gravity-driven  
78 systems remain unclear. Few studies have proposed generic depositional models linking mixed shallow  
79 systems with their potential coeval deep-marine counterparts. However, they mostly rely on subsurface  
80 data (e.g., Moscardelli et al. 2019) and only rarely discuss the impact of changes in structural setting of  
81 the sediment source (e.g., thrust forelimb and backlimb settings) on the deposits (e.g., Čosović et al.  
82 2018).

83 Here, we investigate the type and distribution of shelf-derived sediments, sourced from mixed siliciclastic-  
84 carbonate systems (*i.e.*, compositional systems (*sensu* Chiarella et al. (2017))) that developed above the  
85 asymmetrical basin-bounding structures (*e.g.* thrusts and or fault-growing folds) of an actively deforming,  
86 confined intra-slope basin. We examine the vertical and lateral stratigraphic variability of the shelf-derived  
87 gravity-driven deposits, at outcrop-scale, from the Whareama trench-slope basin, in the emerged  
88 southern portion of the Hikurangi subduction margin, North Island of New Zealand. The deposits are of  
89 Middle Miocene age and incorporate similar material (nature and size of the clasts), albeit transported  
90 through a variety of sediment gravity flows, thereby suggesting different source, delivery and geotectonic  
91 settings across the same confined intra-slope basin.

92 Our study aims to determine how variations in the structural style (thrust forelimb and backlimb) and  
93 tectonic activity (renewal or quiescence) influence the generation of contrasting mixed siliciclastic-  
94 carbonate shelf-derived gravity-driven systems and deposits across a confined, intra-slope basin (*e.g.*,  
95 trench-slope basin).

96 Specific objectives are to:

- 97 • Document the nature, architecture, and size of the different styles of shelf-derived sedimentation  
98 that can occur in a confined intra-slope basin, through an integrated interpretation of  
99 photogrammetry, fieldwork and taphonomic data.
- 100 • Gain new understandings on the controls on sediment transfer and distribution beyond the edges  
101 of mixed siliciclastic-carbonate shelves located upon and around the forelimb and backlimb of  
102 actively growing asymmetrical structures.
- 103 • Develop generic depositional models of the distribution and styles of shelf-derived systems and  
104 subsequent deposits in tectonically active settings.
- 105 • Determine the implications for the stratigraphic prediction of deep-marine, shelf-derived gravity-  
106 driven systems and, of shelf-derived mass-wasting systems along active margins.

## 107 **3. GEOLOGICAL SETTING**

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### 108 **3.1. Regional setting**

109 The study area is located on the eastern North Island of New Zealand, on the tectonically active Hikurangi  
110 Margin (Figure 1). The Hikurangi subduction wedge (*sensu* Bailleul et al. 2013) formed in response to  
111 the Uppermost Oligocene-to-Recent westward subduction of the Pacific Plate beneath the Australian

112 Plate (Ballance 1976; Spörl 1980; Pettinga 1982; Chanier and Ferrière 1991; Field et al. 1997; Nicol et  
113 al. 2007). Bounded by the Hikurangi Trench to the east and the Forearc Basin to the west, the Hikurangi  
114 subduction wedge comprises a succession of elongated trench-parallel sedimentary basins (*i.e.*, trench-  
115 slope basins) separated by tectonically active structural ridges (Chanier and Ferrière 1991; Lewis and  
116 Pettinga 1993; Bailleul et al. 2013; McArthur et al. 2019).

117 Since the onset of subduction, the Hikurangi subduction wedge underwent a polyphase tectonic history,  
118 comprising three main tectonic periods (Figure 2 and references within). Complex stratigraphic  
119 architectures and geometries, such as diachronous sedimentation patterns, syn-sedimentary  
120 deformations and discontinuities, are characteristic of the Miocene-to-Recent trench-slope basin fills and  
121 attest of the close interplay between tectonics and sedimentation (Neef 1992; Neef 1999; Bailleul et al.  
122 2007; Bailleul et al. 2013; Burgreen and Graham 2014; McArthur et al. 2019; McArthur et al. 2021). Deep-  
123 marine sedimentation dominates their fill and includes mass-transport deposits (MTDs), turbidites and  
124 extensive hemipelagic mudstones; yet carbonate, biogenic and shallow-marine fringe deposits can also  
125 be found sporadically (Figure 2) (Field et al. 1997; Lee and Begg 2002; Bailleul et al. 2007; Bland et al.  
126 2015; McArthur and McCaffrey 2019). These basin fills either conformably or unconformably overlie the  
127 Cretaceous to Paleogene pre-subduction basement, mainly composed of the Lower Cretaceous Torlesse  
128 greywackes and the Upper Cretaceous to Oligocene detrital to pelagic series (Spörl 1980; Bradshaw  
129 1989; Chanier and Ferrière 1991; Field et al. 1997; Lee and Begg 2002).

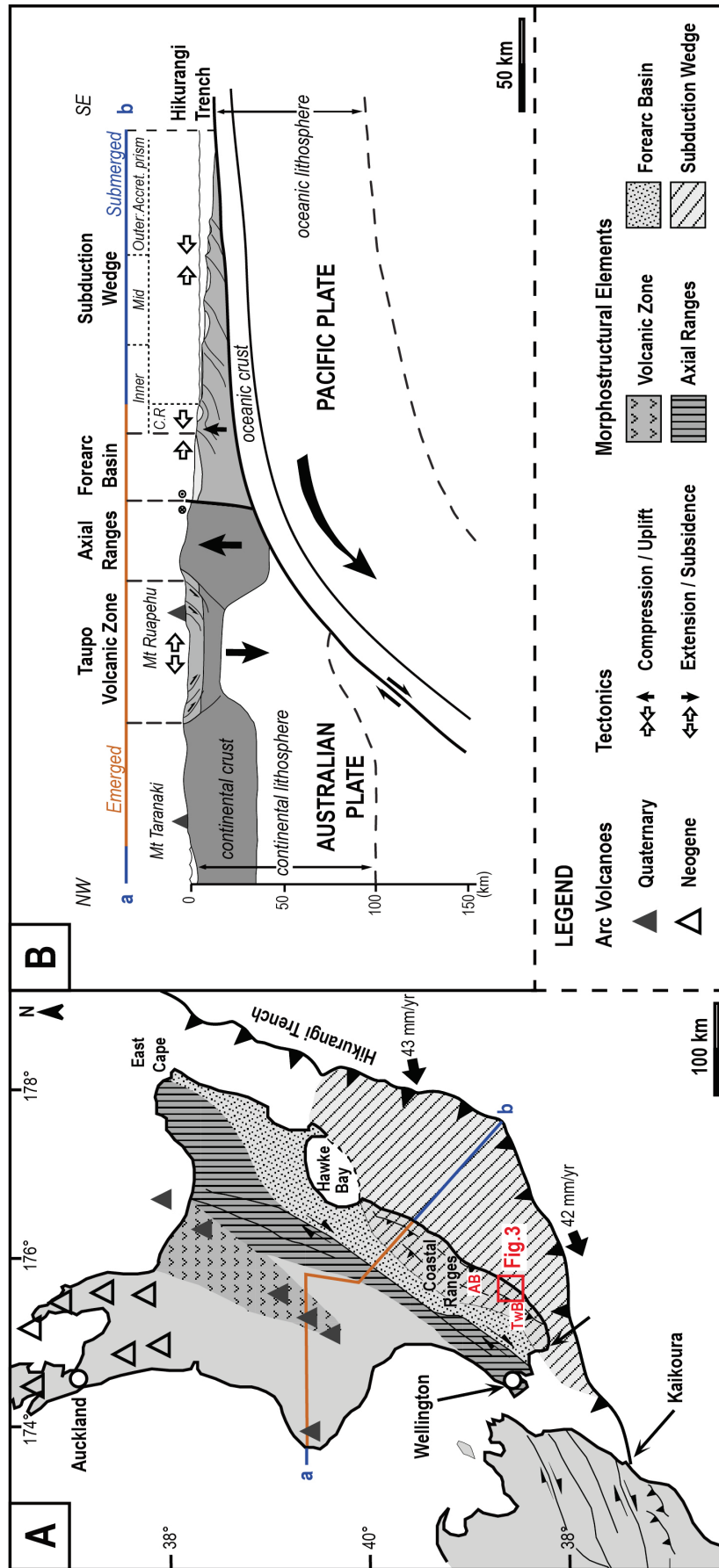


Figure 1: (A): Major subduction-related morphostructural features of the active Hikurangi Margin, north Island of New Zealand. Black arrows show present-day relative plate motion between Pacific and Australian Plates from Beavan et al. (2002). (AB – Akito Basin; TwB – Te Wharau Basin). See (B) for the a – b general cross-section of the Hikurangi subduction complex (C.R – Coastal Ranges). Modified after Chanier et al. (1999), Baillieu et al. (2007) and Baillieu et al. (2013).

130

### 3.2. Southern Coastal Ranges

131 The study area lies within the exhumed inner portion of the subduction wedge (e.g., emergent trench-  
132 slope break) that forms the southern Coastal Ranges of the eastern North Island of New Zealand. Here,  
133 we focus on the tectonic settings at the end of the first compressional phase (Figure 2; Middle Miocene,  
134 Clifdenian – Lillburnian (Langhian to early Serravallian, 15.9 – 13.05 Ma)), where both the inception and  
135 development of the trench-slope basins occurred (Chanier and Ferrière 1991; Bailleul et al. 2013; Malie  
136 et al. 2017).

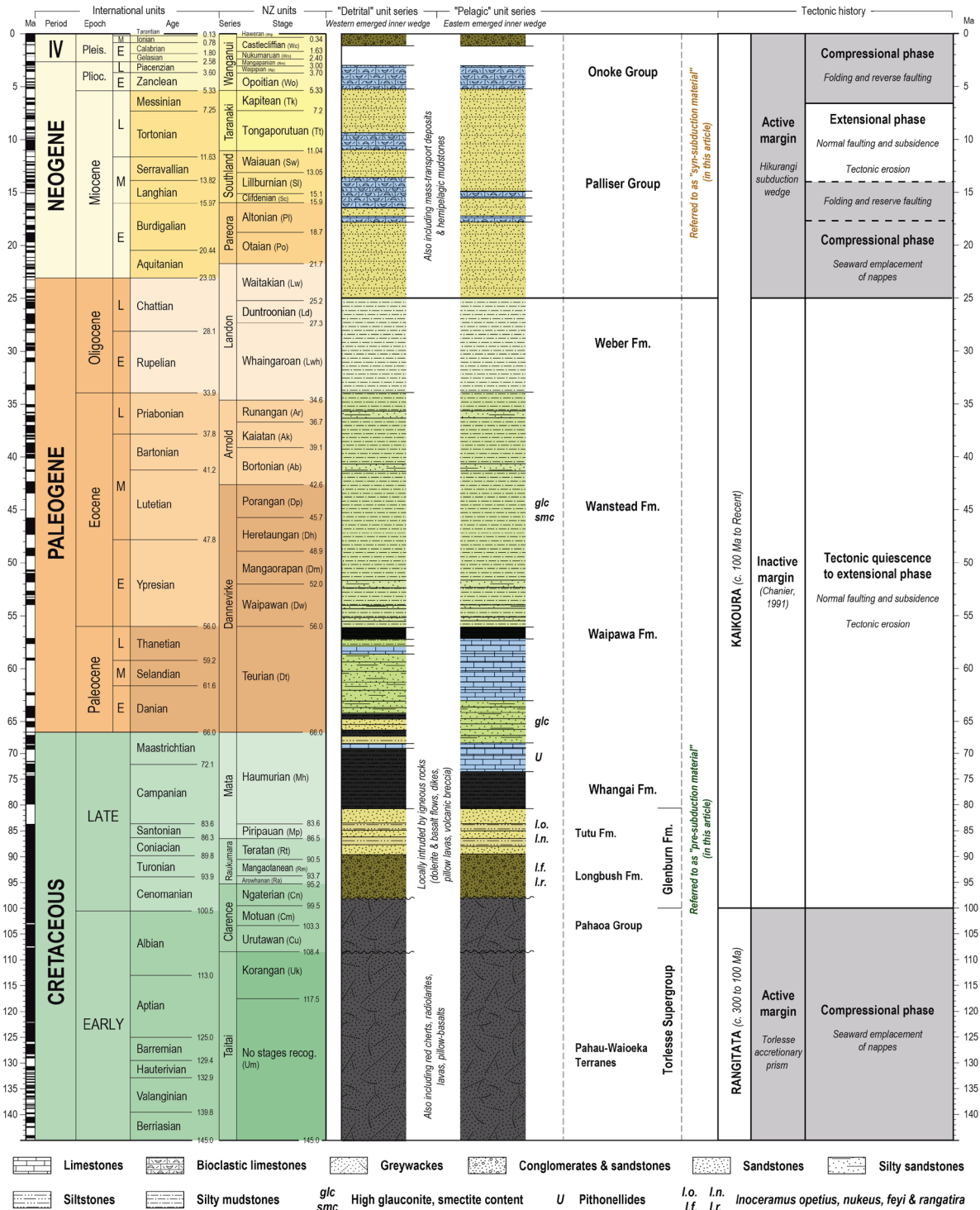
137 During this time period, compressional tectonics resulted in uplift of some of the fold and thrust basin-  
138 bounding structures and more generally of the margin (Crundwell 1987; Bailleul et al. 2013). The induced  
139 changes in depositional environments (e.g., neritic conditions) atop the structures controlled the  
140 development of shelfal environments, and, in particular, mixed siliciclastic-carbonate systems (i.e.,  
141 compositional systems (*sensu* Chiarella et al. (2017)) at shallower waters whilst gravity-systems  
142 dominated elsewhere at deeper waters (Crundwell 1987; Chanier and Ferrière 1991; Neef 1992; Lee and  
143 Begg 2002; Bailleul et al. 2007; Bailleul et al. 2013; McArthur et al. 2019; McArthur and McCaffrey 2019;  
144 Crisóstomo-Figueroa et al. 2020; McArthur et al. 2021).

145 Numerous mixed siliciclastic-carbonate systems co-existing during this period formed a regional shelf  
146 domain throughout the south-western portion of Hikurangi Margin (Crundwell 1987; Chanier 1991;  
147 Bailleul et al. 2007; Bailleul et al. 2013; Bailleul et al. 2015; Claussmann et al. 2021). The shelves  
148 developed above substantially different substrata inherited from local tectonics, and they either  
149 unconformably overlie Lower to Middle Miocene syn-subduction sedimentary rocks or Cretaceous to  
150 Paleogene pre-subduction basement. Their depositional settings preferentially suggest narrow,  
151 continent-attached shelfal systems receiving regular terrigenous inputs from the hinterland (Bailleul et al.  
152 2015; Claussmann et al. 2021). However, isolated, continent-detached platform systems may also have  
153 locally occurred on some of the actively growing structures (see (Caron et al. 2004; Caron et al. 2021)  
154 for examples along the same margin, yet at different geological times).

155 Claussmann et al. (2021) previously demonstrated that short-lived periods (ca. 1 to 2 Ma) of sustained  
156 (with or without phases of quiescence) compressional tectonics favor the propagation of the folds and  
157 thrusts, and thus the expansion of abrupt, unstable areas close to these structurally-controlled shelf-  
158 margins. Repeated uplift and slope oversteepening of these areas eventually lead to recurrent shelf  
159 collapses, sourcing shelf-derived sediments (e.g., mass-wasting products) into deep water. The  
160 associated deposits result from sediment gravity flows, such as cohesive debris flows or mudflows, and



161 rework a substantial amount of shelf-derived material, incorporating sediments and fossils from neritic  
 162 shelfal environments. Several occurrences of such shelf-derived mass-wasting events and related  
 163 deposits of Middle Miocene age were identified in the southern Coastal Ranges cropping out across  
 164 several trench-slope basins (*i.e.*, Whareama, Te Wharau and Akitio Basins) (Figure 1; Figure 2; Figure  
 165 3).



166

167 Figure 2: [previous page] Chronostratigraphic chart for the southern, emerged portion of the Hikurangi subduction wedge  
168 adapted from [Claussmann et al. \(2021\)](#). Lithostratigraphy details adapted from [Chanier et al. \(1990\)](#), [Chanier \(1991\)](#), [Chanier](#)  
169 [and Ferrière \(1991\)](#), [Field et al. \(1997\)](#), [Lee and Begg \(2002\)](#) and [Bland et al. \(2015\)](#) and detailing the pre- and syn-Hikurangi  
170 subduction series. Regional tectonism adapted from [Chanier et al. \(1999\)](#), [Bailleul et al. \(2013\)](#) and [Malie et al. \(2017\)](#). New  
171 Zealand stages after [Raine et al. \(2015\)](#) showing the equivalence with the international stages.

### 172 **3.3. The Whareama Basin**

173 The Whareama Basin is narrow (two to six kilometers wide), elongated (50 kilometers long) and trench-  
174 parallel (NE-SW) ([Johnston 1980](#); [Chanier 1991](#)). It is bounded by pre-Miocene basement ridges that are  
175 controlled by the Adams-Tinui Fault complex and Pukeroro Fault to the west (*i.e.*, landward basin margins,  
176 Figure 3) and the Flat Point-Whakataki Fault complex to the east (*i.e.*, seaward basin margin, Figure 3).  
177 The basin's development began at the onset of subduction and thus records gravity-driven deposits  
178 dating from the earliest Miocene. It formed on the back of the Glenburn Nappe, a trenchward-  
179 advancing thrust sheet composed of Cretaceous to Eocene pre-subduction series that are either  
180 unconformably or conformably overlain by syn-subduction deposits ([Chanier 1991](#); [Chanier and Ferrière](#)  
181 [1991](#)).

182 The study area is located on the eastern limb of the Whareama Basin, with beds strongly dipping towards  
183 the west (ca. 40°). Nearby, well-exposed Middle Miocene shelf-derived MTDs were previously described  
184 and attributed to result from the collapses of coevally developing narrow, continent-attached shelf-  
185 margins located to the north-west (Figure 3; Sefton Hills s-1 and s-2 sections) ([Claussmann et al. 2021](#)).

## 186 **4. DATA AND METHODS**

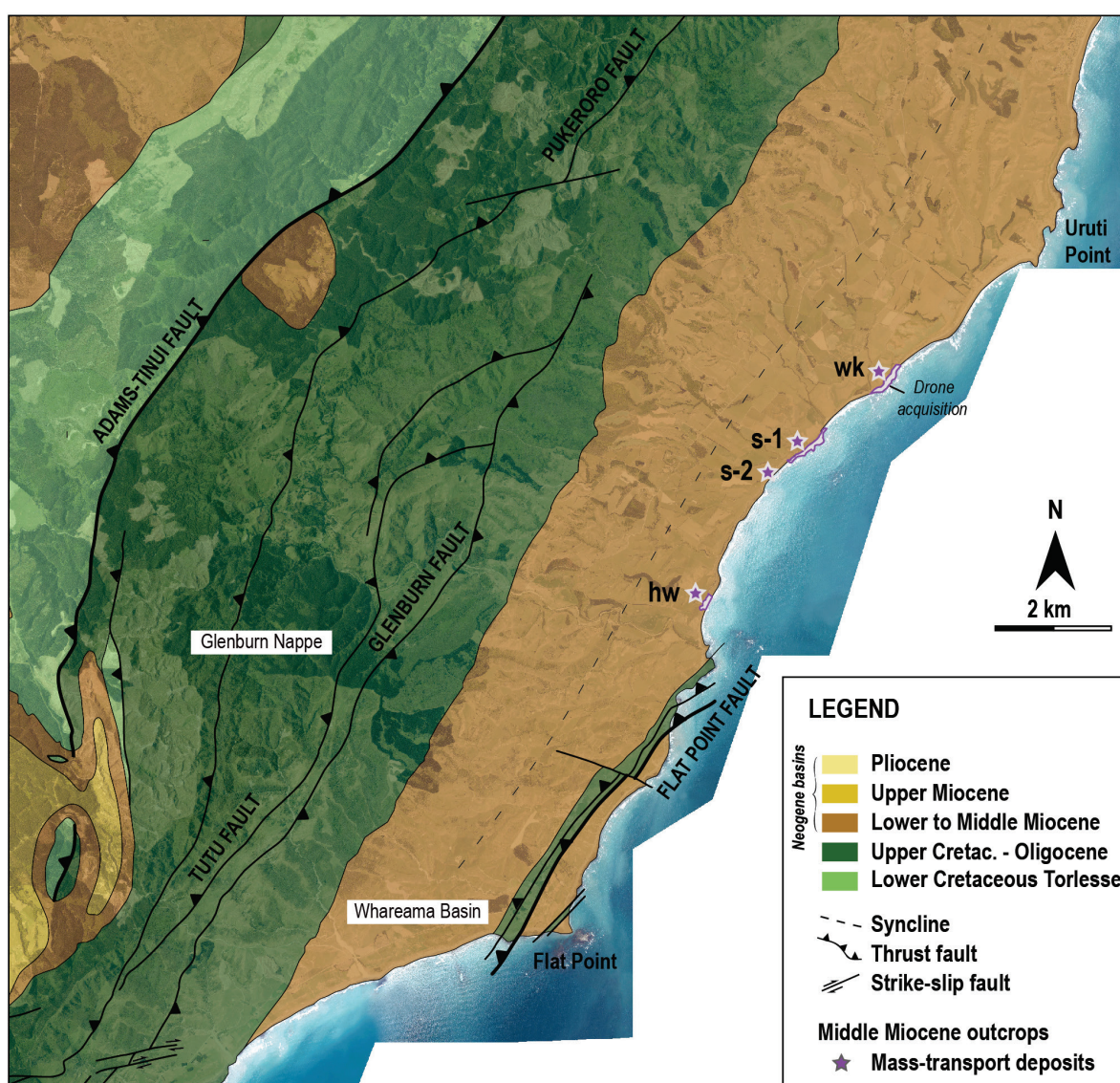
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187 This outcrop-based study integrates traditional and digital field data acquired during three field campaigns  
188 (2018, 2019, 2020) in the southern portion of the Hikurangi subduction wedge (Coastal Ranges, eastern  
189 North Island of New Zealand), and focuses on the shelf-derived gravity-driven deposits of the Waikaraka  
190 and Homewood localities (Figure 3; wk and hw sections).

191 Fieldwork consisted of logging sedimentological profiles (total of 312 meters at 1:50 bed-scale),  
192 measuring structural (*e.g.*, ductile deformation analysis) and stratigraphic features (*e.g.*, bedding,  
193 paleocurrent) as well as describing each shelf-derived flow deposit in a standardized manner (facies and  
194 geometries).



195 Owing to the abundant macrofossils contained in the deposits, occurring both as whole skeletons and  
 196 fragmented remains, five taphonomic analyses were performed on approximately one square meter  
 197 areas for both the outcrops (*i.e.*, four at Waikaraka and one at Homewood). These analyses were  
 198 restricted to the coarsest fraction (larger than five millimeters) of death assemblages contained in the  
 199 deposits. Three categories of skeleton damages, namely fragmentation, abrasion and bioerosion, were  
 200 described visually using the graded classification scale presented in Appendix 1, complemented by Caron  
 201 (2011) and Caron et al. (2019).



202  
 203 Figure 3: Satellite map from World Imagery (ESRI), and onshore geological map from Chanier (1991) of the Whareama Basin.  
 204 Location of the drone acquisition and related 3D outcrop models presented in this study = wk = Waikaraka section and hw =  
 205 Homewood section. See Claussmann et al. (2021) for the s-1 and s-2: Sefton Hills section and 3D outcrop model.

206 Seven samples were collected to supplement the Fossil Record Electronic Database (<https://fred.org.nz/>)  
207 since both areas are currently mapped as Quaternary beach deposits on the 1:250 000 Geological Map  
208 of New Zealand (QMAP) of Wairarapa (<https://www.gns.cri.nz/>). The micro- and macropaleontological  
209 analyses were conducted by GNS New Zealand and allowed to determine the age of the sedimentary  
210 units and related depositional paleobathymetries (Appendix 2; Appendix 3).

211 Digital fieldwork involved the acquisition of high-resolution aerial images (1,900 at Waikaraka and 690 at  
212 Homewood) using a DJI Phantom 4 Pro drone, supplemented by a number of Ground Control Points  
213 (GCP) collected with a Trimble GeoExplorer 2008 differential global positioning system tool (DGPS) and  
214 a Trimble Tempest antenna positioned at two meters above the measured point. Two georeferenced 3D  
215 outcrop models were then created in the form of high-resolution triangulated mesh textured with the  
216 photographs.

## 217 5. FACIES AND ARCHITECTURAL ELEMENTS

---

### 218 5.1. Facies and architectural schemes

219 Integration of the data allowed an analysis and comparison of the stratigraphic architectures and facies  
220 organization within both outcrops.

221 A total of 14 lithofacies were recognized across the two outcrops (summarized in Table 1 and illustrated  
222 in Figure 4, Figure 5, Figure 6 and Figure 7) and classified according to their dominant lithology, primary  
223 sedimentary features and interpreted in terms of flow processes using models from (Bouma 1962; Nardin  
224 et al. 1979; Lowe 1982; Postma 1986; Kneller 1995; Mulder and Alexander 2001).

225 These comprise massive silty mudstones (**MDST**); tabular, very thin- to medium-bedded (**LDTC-a**) or  
226 medium- to thick-bedded (**LDTC-b**) sandstones with mudstone caps; lenticular, medium to very-thick  
227 bedded sandstones (**HDTC**); structured (**SST-a**) or structureless (**SST-b**) sandstones; pebbly sandstones  
228 (**sHDTC-a**); bioclastic grits (**sHDTC-b**); lenticular, medium to very-thick bedded gravel sandstones with  
229 mudstone caps (**gsHDTC**); organized to disorganized clast-supported (respectively **gHDTC**, **CF**) and  
230 matrix-support (**DF**) conglomerates; as well as deformed (**SL-a**) or undeformed (**SL-b**) mass of  
231 sediments.



Deposit	Dominant lithology	Lithofacies code	Thickness (cm)	Lithofacies description	Classification and process interpretation
Turbidite	Mudstone	<b>MDST</b> Massive, siltstone to silty mudstone	1 to 60 (HMD) nul (WKS)	Siltstone to silty mudstone. Mainly massive, rare parallel lamination visible. Ungraded to (normally) graded. Variable degree of bioturbation, tends to be highly bioturbated.	(Td)-Te Low-density turbidity currents and hemipelagic suspension settling
	Sandstone to Mudstone	<b>LDTC-a</b> Tabular, very thin-to medium-bedded sandstone with mudstone cap	1 to 20 (HMD) 2 to 20 (WKS)	Fine- to medium-grained. Parallel laminations (Tb), passing into climbing ripples (Tc) with sometimes dewatering or soft sedimentation deformation structures installed above. Sometimes only developing the parallel laminations or sometimes directly developing the climbing ripples. Laminations can be highlighted by organic-rich, carbonaceous debris and shell ashes. Common bioturbation.	Tb-e Tb-Te Tc-Te Low-density turbidity currents
		<b>LDTC-b</b> Tabular, medium to thick-bedded sandstone with mudstone cap	5 to 100 (HMD) 10 to 100 (WKS)	Fine- to coarse-grained. If present, medium- to very coarse-grained Ta at the base of the thicker beds, commonly massive and structureless, with sand- to granule-(up to pebble-) grade clasts (lithoclasts, bioclasts and rip-up mudstone clasts), sometimes displaying sole marks such as flute casts. Fine- to medium-grained, well-developed parallel laminations (Tb), climbing ripples (Tc) with often dewatering or soft sedimentation deformation structures installed above. Common amalgamation. Lamination often highlighted by organic-rich, carbonaceous debris and shell ashes. Common bioturbation.	Ta-e Low-density turbidity currents
		<b>HDTC</b> Lenticular, medium to very thick-bedded sandstone with mudstone cap	5 to 20 (HMD) 5 to 200 (WKS)	Fine- to coarse-grained. Lenticular, medium- to very-coarse-grained base, commonly massive and structureless, with sand- to granule-(up to pebble-) grade clasts (lithoclasts, bioclasts and (elongated) rip-up mudstone clasts), overlain by well-developed parallel or cross-laminations in the same material, overall normally graded. Followed by fine- to medium-grained, well-developed parallel laminations (Tb), climbing ripples (Tc) with often dewatering or soft sedimentation deformation structures installed above. Lamination often highlighted by organic-rich, carbonaceous debris and shell ashes. Common bioturbation. HMD only presents the bottom part.	S-Ta-e Sandy high- to low-density turbidity currents
	Sandstone	<b>SST-a</b> Structured sandstone, above gHDTC, CF, DF or MF	5 to 20 (HMD) 15 to 100 (WKS)	Fine- to medium-grained. Parallel, sometimes passing into faint climbing ripples and or dewatering or soft sedimentation deformation structures, usually normally graded. Common scattered shell ashes; organic-rich, carbonaceous debris, and or granule- (to pebble-) grade lithoclasts. Mudstone cap commonly absent.	Tb-Tc Turbulent cloud, deposition from traction/traction-plus-fallout
		<b>SST-b</b> Structureless sandstone, above gHDTC, CF, DF or MF	5 to 30 (HMD) 15 to 100 (WKS)	Fine- to coarse-grained. Structureless, sometimes dewatering structures. Ungraded or (normally) graded. Common scattered shell ashes; organic-rich, carbonaceous debris, and or granule- (to pebble-) grade lithoclasts. Mudstone cap commonly absent.	S Turbulent cloud, rapid suspension fall-out
	Conglomerate to Sandstone	<b>sHDTC-a</b> Pebbly sandstone	5 to 30 (HMD) nul (WKS)	Fine- to coarse-grained. Structureless, or crude parallel laminations, commonly normally graded. Presenting extraformational clasts and rip-up mudstone clasts scattered or organized. Common mudstone cap.	S-Te Sandy high-density turbidity currents
		<b>sHDTC-b</b> Dis- to organized clast-supported, bioclastic	1 to 20 (HMD) nul (WKS)	Clast-supported, siltstones to silty mudstones with ≤90% of granule- to pebble-grade, extra- and intraformational clasts. Ungraded, sometimes normally graded laterally. Highly bioclastic with possible shell ashes organization. Frequent rip-up mudstone clasts in the normally graded sections. Often overlying sand with parallel laminations and common load structures, and either overlain by structureless sand and or mudstone cap, rarely bioturbated.	R?-S/Ta?-Te Cohesionless debris flow or bed-load transport underneath gravelly to sandy high-density turbidity currents
		<b>gsHDTC</b> Organized gravel to fine-grained sandstone with mudstone cap	nul (HMD) 50 to 300 (WKS)	Combination of facies gHDTC (sometimes even facies CF) at the base, overlain by facies HDTC. Clast-supported interval particularly rich in shell ashes and debris content, mostly composing the binding link between the lithoclasts; frequent rip-up mudstone clasts; very frequent boulder-graded clasts marking R2-R3 transition; low-angle, lenticular, slightly erosive base. Common amalgamation in HDTC interval.	(R2)-R3-S/Ta-e Gravelly to sandy (bipartite) high-density turbidity currents: bed-load transport underneath high-density turbidity currents
	Conglomerate	<b>gHDTC</b> Organized clast supported	10 to 100 (HMD) 10 to 200 (WKS)	Clast-supported, siltstone to silty mudstone with ≤90% of granule- to boulder-grade, extra- and intraformational clasts. Highly bioclastic when of average granule-grade clast size, with possible shell ashes organization. Graded (inverse to normally graded, or only normally graded). Occasional clast imbrication / organization and load structures. Rare to locally frequent rip-up mudstone clasts.	R2-R3 R3 Gravelly (bipartite) high-density turbidity currents: bed-load transport underneath high-density turbidity currents
Debrite		<b>CF</b> Disorganized clast-supported	10 to 200 (HMD) 20 to 400 (WKS)	Clast-supported, siltstone to silty mudstone with ≤90% of granule- to boulder-grade extra- and intraformational clasts. Highly bioclastic when of average granule-grade clast size. Ungraded, disorganized. Common load structures.	R1 Cohesionless debris flow or bed-load transport underneath gravelly high-density turbidity currents
	<b>DF</b> Disorganized matrix-supported	150 to 400 (HMD) 100 to 400 (WKS)	Matrix-supported, siltstone to silty mudstone with varying quantity of granule- to boulder-grade extra- and intraformational clasts. Ungraded, disorganized. Common recumbent folds, shear and load structures. Traction carpet possible.	/ Cohesive debris flow to cohesionless, sheared debris flow when comprising facies CF at base, top or edges	
Slide-Slump	Alternating conglomerate, sandstone and or mudstone	<b>SL-a</b> Deformed mass of sediments	100 to 2000 (HMD) nul (WKS)	Contorted, remobilized facies characterized by recumbent folds. Commonly sandy to silty mudstone background sediments.	/ Slump: coherent mass of sediment that moves along a glide plane, resulting in significant internal deformation
		<b>SL-b</b> Undeformed mass of sediments	100 to 2000 (HMD) 100 to 200 (WKS)	Coherent, remobilized facies displaying sharp truncations, slightly concave-up geometries with downlap.	/ Slide scar / scour: initiation of movement of a coherent mass of sediment along a glide plane, no internal deformation

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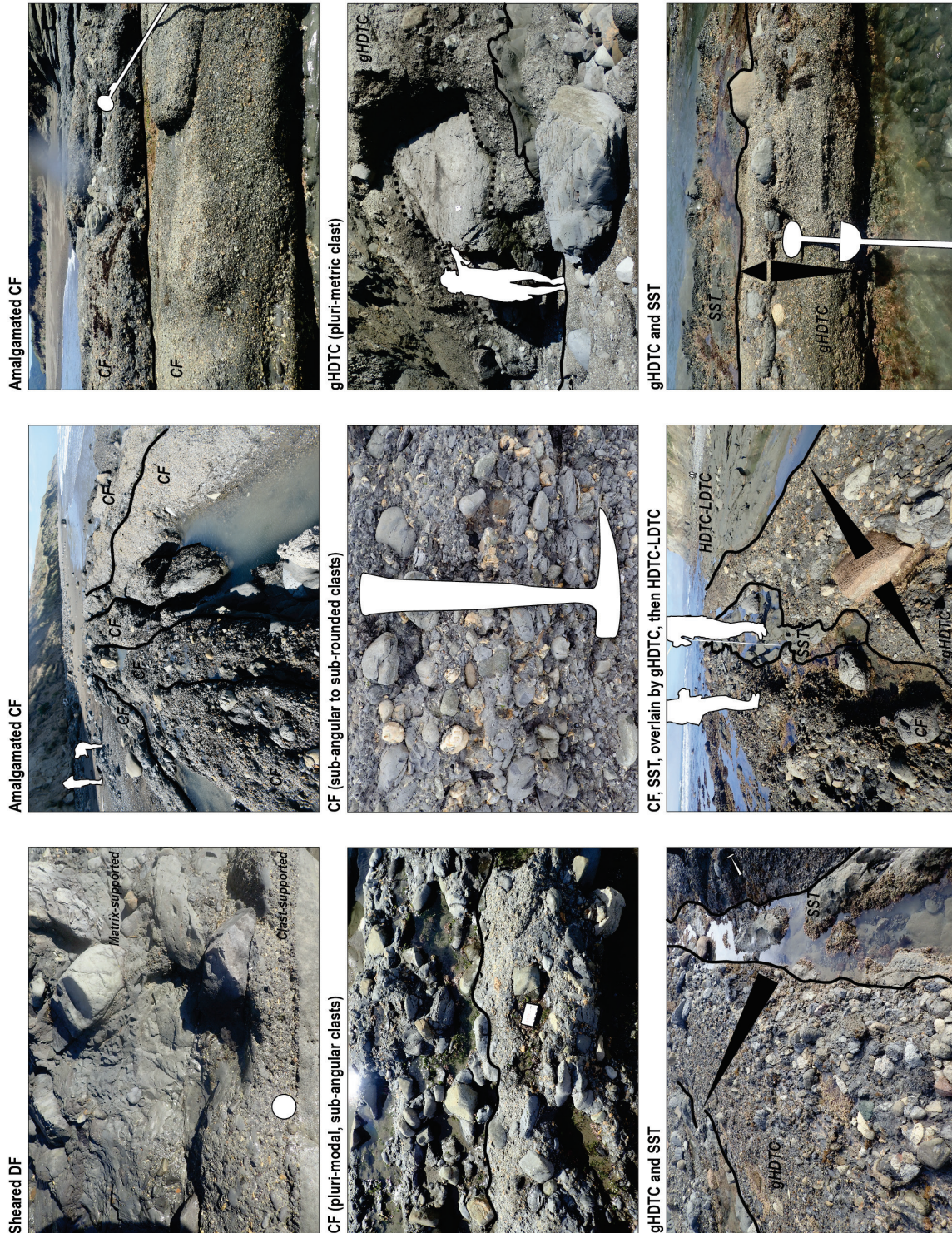
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Table 1: Lithofacies encountered at the Waikaraka (WKS) and Homewood (HMD) outcrops. See Figure 4 and Figure 5 for representative photographs of each lithofacies at Waikaraka; and Figure 6 and Figure 7 at Homewood. Note that the 'Classification' column refers to (Bouma 1962) and (Lowe 1982), and that extraformational clasts include pre- and syn-subduction lithoclasts, and syn-subduction bioclasts. Intraformational clasts include syn-subduction lithoclasts.



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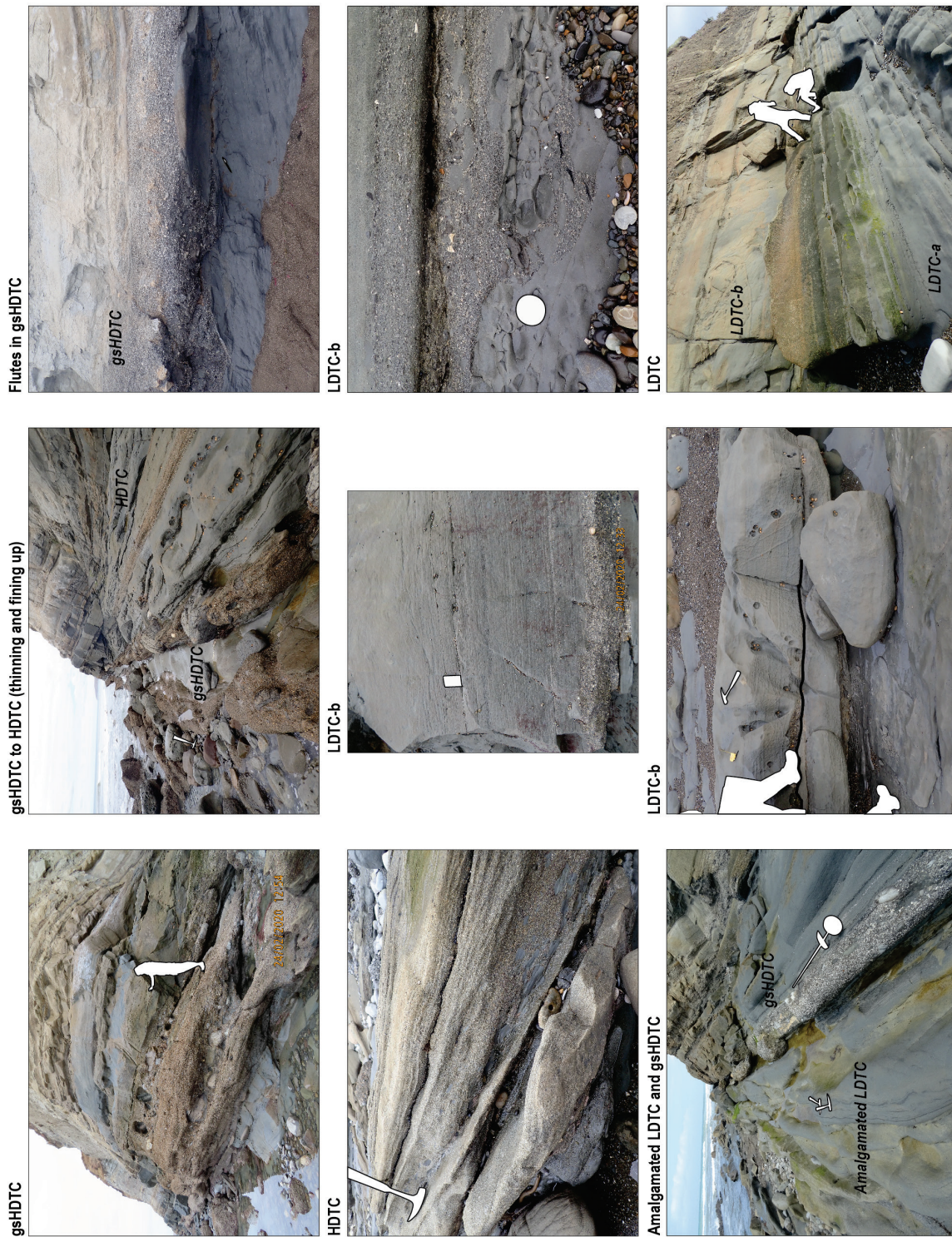
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Figure 4: Representative photographs of the lithofacies summarized in Table 1 and recognized at Waikaraka. Scales: human size (~1.70-1.75 meters); Jacob's staff (1.5 meter); hammer (33 centimeters); card (8.5 centimeters long); coin (2.3 centimeters).



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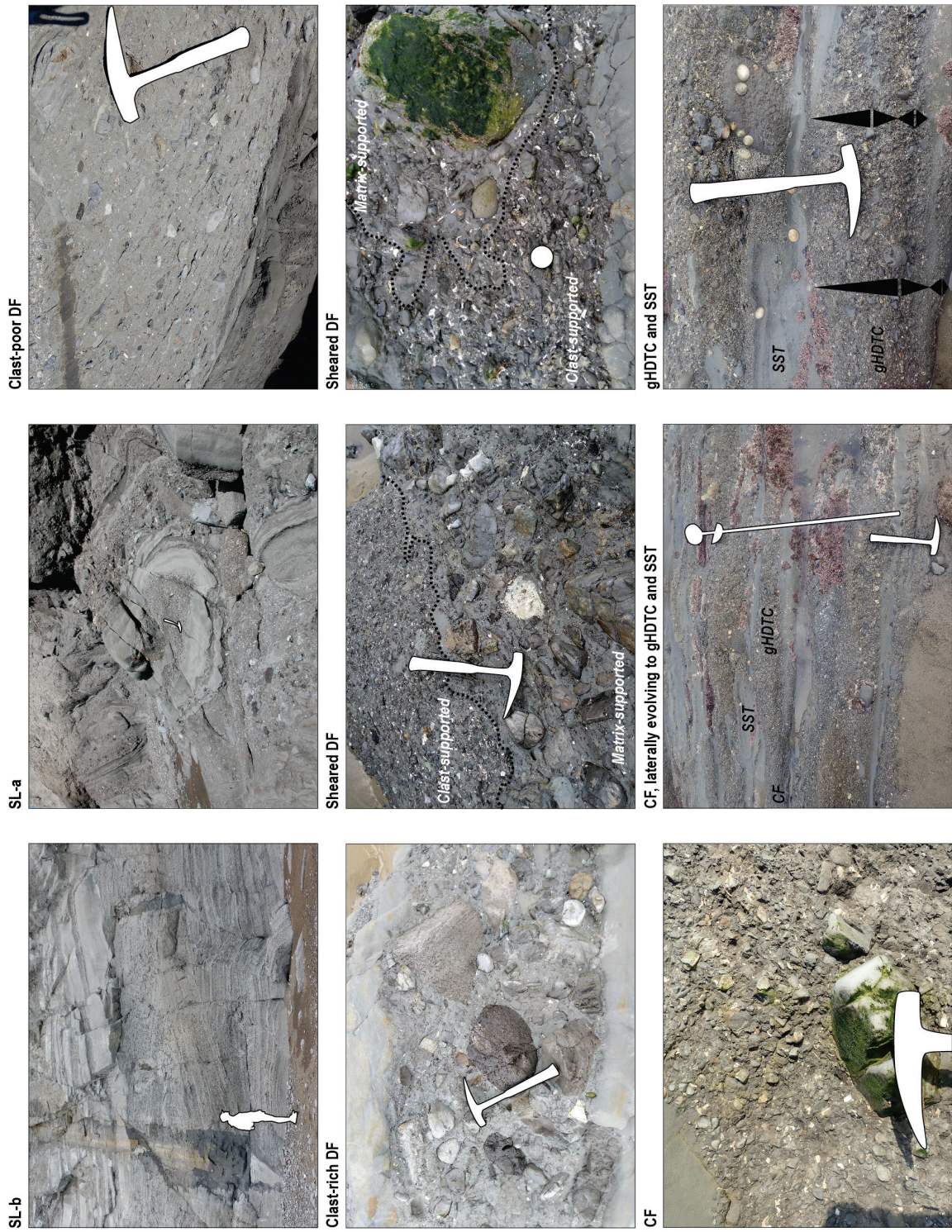
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Figure 5: Representative photographs of the lithofacies summarized in Table 1 and recognized at Waikaraka. Scales: human size (~1.70 meters); Jacob's staff (1.5 meter); hammer (33 centimeters); card (8.5 centimeters long); coin (2.5 centimeters).

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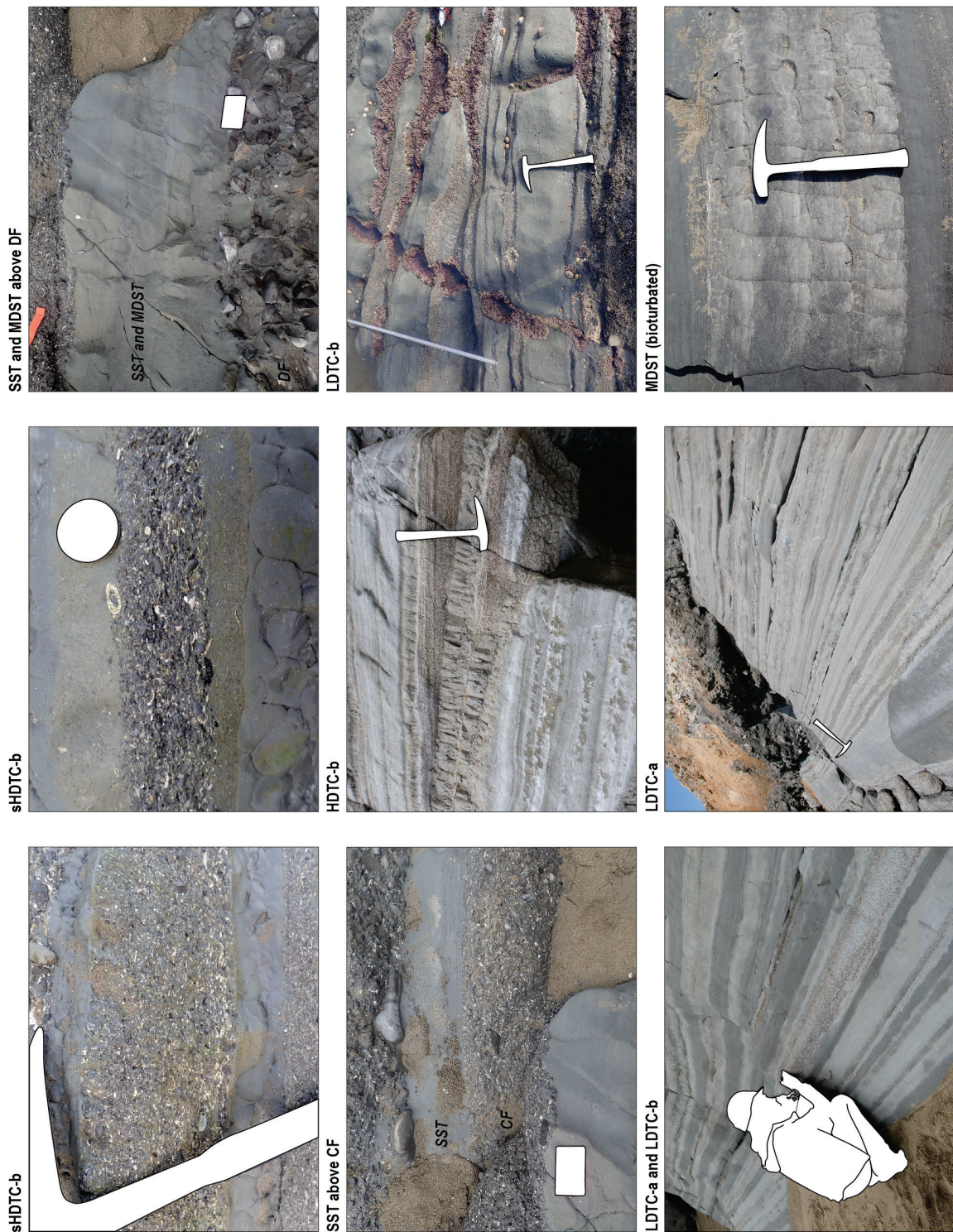


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Figure 6: Representative photographs of the lithofacies summarized in Table 1 and recognized at Homewood. Scales: human size (~1.70 meters); Jacob's staff (1.5 meter); hammer (33 centimeters); card (8.5 centimeters long); coin (2.3 centimeters).

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Figure 7: Representative photographs of the lithofacies summarized in Table 1 and recognized at Homewood. Scales: human size (~1.70 meters); Jacob's staff (1.5 meter); hammer (33 centimeters); card (8.5 centimeters long); coin (2.3 centimeters).

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They were then grouped into three main facies associations, which are related to main depositional systems (**Fa1g**, **Fa2c**, **Fa3p**) (Figure 8; Figure 9). The nomenclature we adopted follows the one defined by Bailleul et al. (2007), Bailleul et al. (2013) and Claussmann et al. (2021), in which **Fa1g** refers to

252 depositional lobe settings, **Fa2c** to confined slope channel settings and **Fa3p** to shelf-derived mass-  
253 wasting systems.

254 Each facies association is described and interpreted in sequence below, directly in context of the outcrop  
255 where they were recorded. Figure 10, Figure 11 and Figure 12 illustrate the geometric relationships that  
256 exist between the different lithofacies and their associations as well as some of the sedimentary  
257 structures at different scales. The nature of the extra- and intraformational clasts reworked in the  
258 conglomeratic and coarse sandstone intervals is described separately (see Section 6) and summarized  
259 in Figure 13 and Figure 14.

## 260 **5.2. Confined slope system**

261 The Waikaraka section crops out five kilometers south of the Uruti Point and extends across one kilometer  
262 along the coast (Figure 3). Deposition occurred in the Middle Miocene, Lillburnian to Waiauian (Late  
263 Langhian to Serravallian, 15.1 – 11.63 Ma, Appendix 2) and features the **Fa2c** depositional system, which  
264 can be divided into two main depositional units (**Fa2c-e**, **Fa2c-l**), hereafter described and interpreted in  
265 sequence. These units are bounded by major erosion surfaces. These prominent surfaces can laterally  
266 be traced over several hundreds of meters (>600 meters), exhibit a relief of up to two meters, and  
267 characteristically separate contrasting depositional facies (Figure 8).

### 268 **5.2.1. Confined slope channel: axis to off-axis, early-stage fills (Fa2c-e)**

#### 269 Observations

270 **Fa2c-e** is characterized by crude fining- and thinning-upward series, dominated by coarse-grained facies.  
271 Here, this association crops out over more than 600 meters laterally and is generally between 10 to 50  
272 meters in thickness (Figure 8).

273 **Fa2c-e** mostly comprises amalgamated, very thick-bedded (one to >two meters), disorganized (**CF**) to  
274 organized (**gHDTC**), polymict and clast-supported conglomerates (Figure 4; Figure 10; Figure 11; Table  
275 1). The conglomerates commonly form continuous, broadly lenticular and fairly symmetrical bodies with  
276 low-relief erosional bases, thereby indicating a cross-sectional, rather than longitudinal, view of them.  
277 Their upper topographies are uneven and they are occasionally overlain by thin- to thick-bedded,  
278 structured or structureless sandstones (**SST**), themselves often incised by the overlying conglomerates,  
279 thus forming composite conglomeratic bodies (Figure 4; Figure 8; Figure 11; Table 1). The conglomerates  
280 essentially evolve from (1) being internally disorganized (**CF**), only comprising local internal stratifications,



281 marked by subtle variations in clast grade and alignment, or in matrix concentration (**DF**) to (2) being  
282 internally organized, displaying inverse- to normally- or only normally-graded patterns (**gHDTC**) (Figure  
283 4; Figure 8; Figure 11; Table 1).

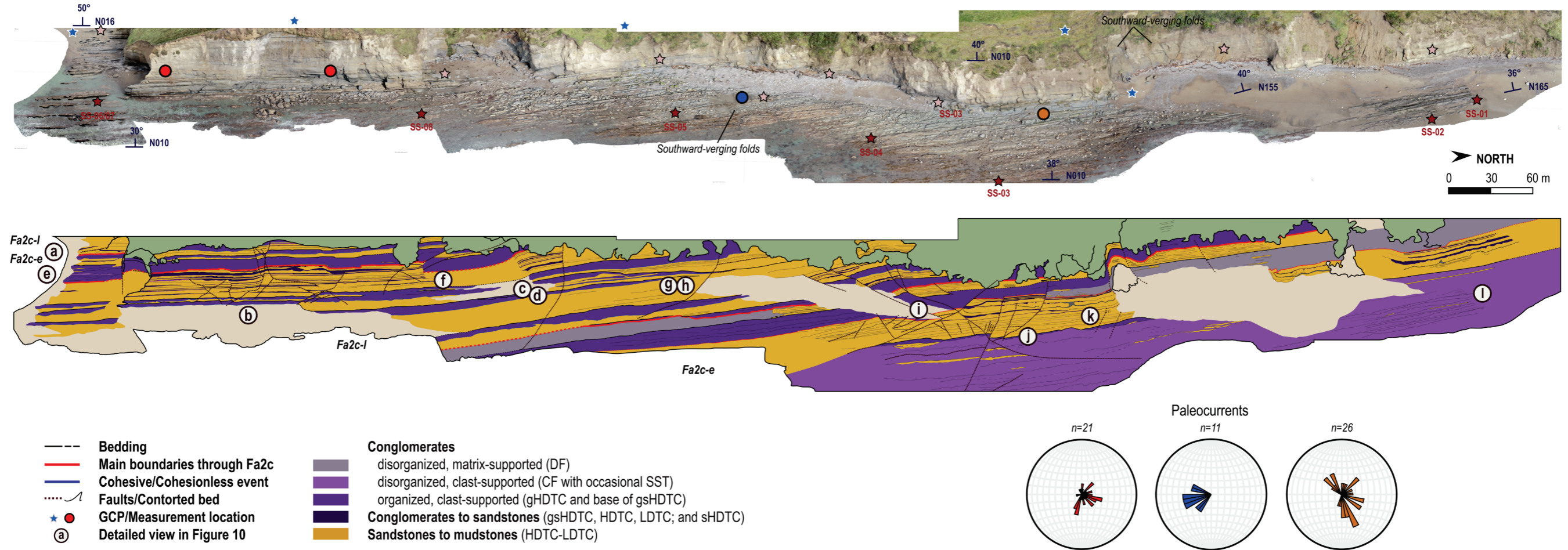
284 The conglomerates are generally overlain by series of laterally continuous, mostly medium- to fine-  
285 grained, thin- to very thick-bedded sandstones, which are either amalgamated or capped by bioturbated  
286 mudstones (**HDTC to LDTC**) (Figure 5; Figure 8; Figure 11j; Table 1). The thicker beds show frequent  
287 grain-size breaks, indicating bed amalgamation (**HDTC to LDTC-b**). Their basal, coarser intervals are  
288 discontinuous, erosive and typically composed of gravel material similar to that of the conglomerates,  
289 which occasionally includes aligned mud clasts, plant debris and some shell fragments (Figure 5, LDTC-  
290 b). The sandstones frequently contain soft-sediment deformation features (e.g., flame, load) in their  
291 structured intervals (Figure 10k), whereas, the mudstone caps, where present, commonly include  
292 *Phycosiphon* sp. and *Chondrites* sp. bioturbations.

293 Four occurrences of **Fa2c-e** were observed at Waikaraka (Figure 8; Figure 11).

294 The first occurrence is exposed at the base of the outcrop. Its basal surface is not visible as it lies below  
295 the sea level. The conglomerates are disorganized (**CF**) (Figure 8; Figure 4; Figure 11) with an average  
296 clast size ranging from granule- and pebble-grades, to pebble- and cobble-grades toward the southward  
297 wavecut platform. Boulders and outsized clasts (deci- to decametric) are frequent (Figure 10l), particularly  
298 to the south and in the disorganized intervals (**CF**). Sub-rounded clasts dominate, but, sub-angular clasts  
299 are common and sometimes restricted to specific intervals (Figure 4; see Section 6). The overlying  
300 sandstone-rich series generally fine and thin-upward, evolving from thick- to medium-bedded, concave-  
301 up sandstones pinching out to the north into nearly-flat bottomed, thinner-bedded sandstones with very  
302 thin- to thin-bedded mudstone caps (**HDTC to LDTC**) (Figure 8; Figure 10j, k; Figure 11). To the south,  
303 amalgamation dominates throughout (Figure 10i; Figure 11). This first occurrence ends with a couple of  
304 laterally continuous, very thick- to medium-bedded sandstones (**LDTC-b**) that are overlain by few thinner-  
305 bedded intervals (**LDTC-a**) and characteristically comprise large-scale fluid escape and soft-sediment  
306 deformation features (Figure 10j, k).

307 The second occurrence directly overlies the first, separated by a basal surface that varies from being  
308 sharp to erosional (> two meters incision) to the south (Figure 8; Figure 10i; Figure 11). The associated  
309 conglomerates are organized (**gHDTC**), mostly showing normally-graded patterns, amalgamated (i.e., no  
310 capping sandstone) and smaller (~one to one and a half meters thick) (Figure 8; Figure 4; Figure 11).

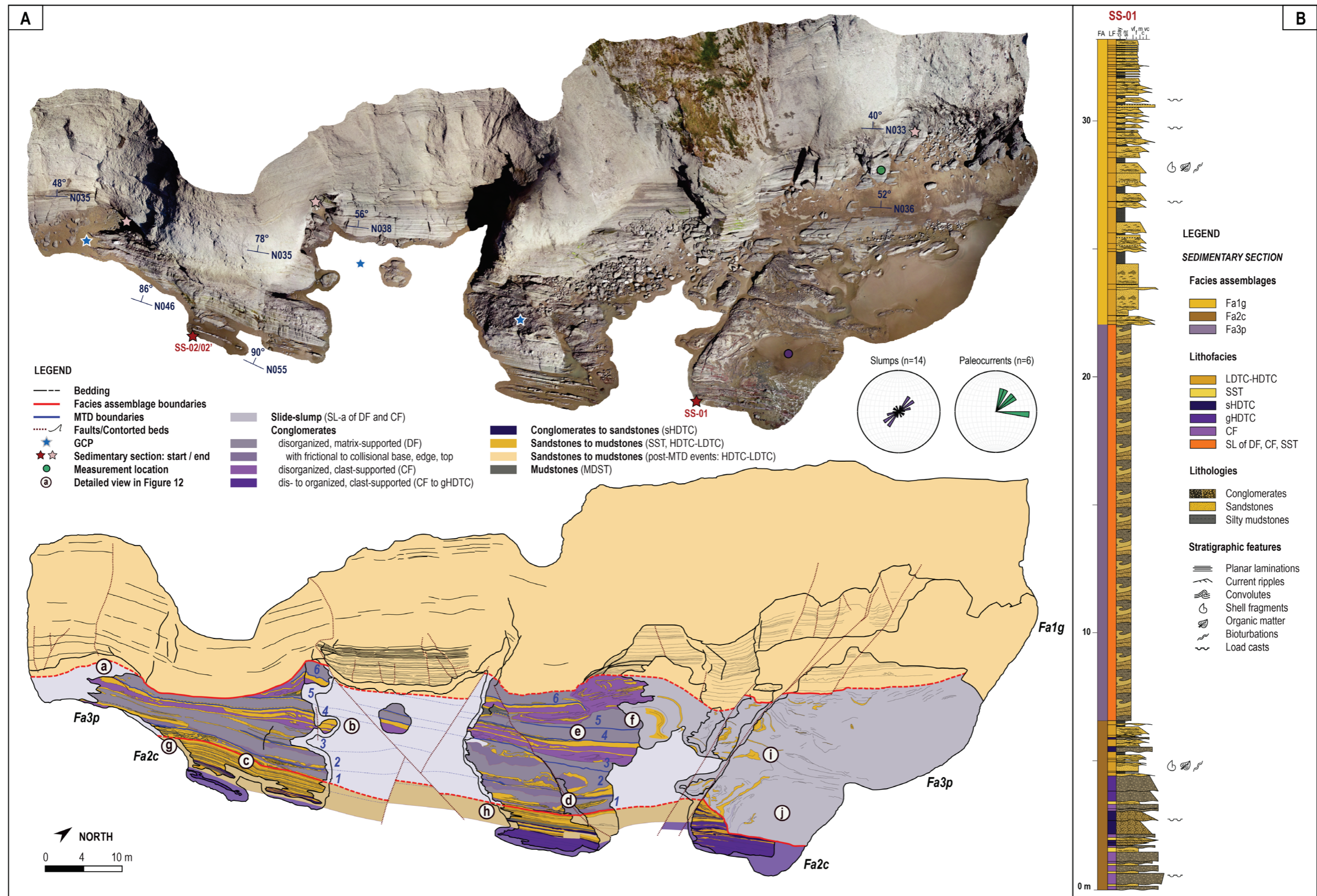
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317 Figure 8: 3D outcrop model (top) and interpretation (bottom) of the Waikaraka coastal outcrop that exposes a confined slope channel system (Fa2c). This system comprises two distinct channel complexes, essentially divided into early-stage confined channel fills (Fa2c-e) and late-stage  
318 confined channel fills (Fa2c-l). The letters refer to remarkable architectural elements that characterize the Waikaraka system: a detailed view and explanation of each of them is available in Figure 10. The colors used in the stereoplots correspond to the location circles presented on the  
319 outcrop model (top). Stereoplots (Schmidt, lower hemisphere) highlight the paleocurrents taken in the turbidite system after back-tilting of bedding planes to initial horizontal position (assuming cylindrical folding).

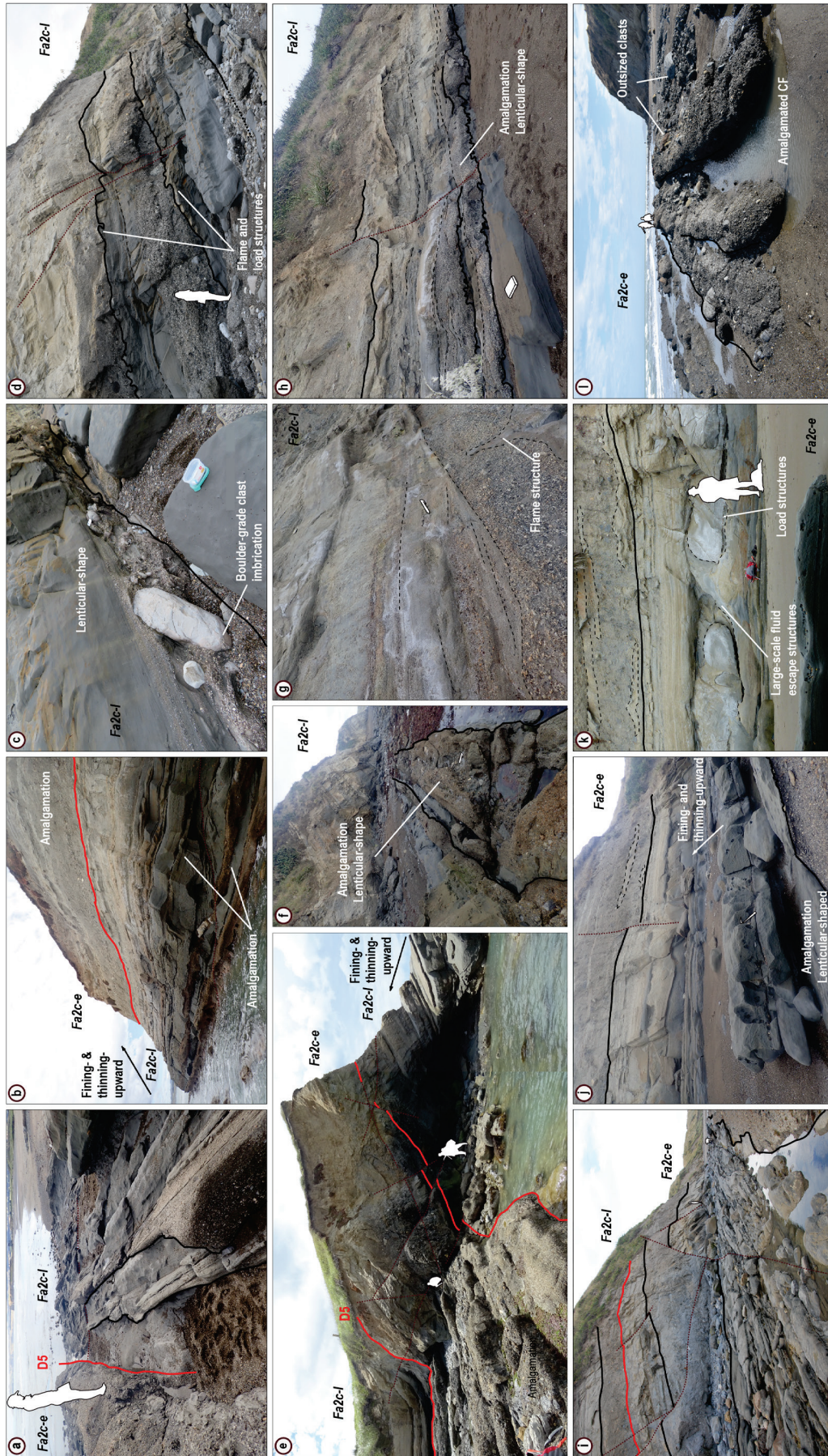




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321 Figure 9: (A): 3D outcrop model (top) and interpretation (bottom) of the Homewood coastal outcrop. This outcrop exposes three distinct gravity-driven systems, from the bottom to the top, a slope channel system (Fa2c), a shelf-derived mass-wasting system (Fa3p) and a distal depositional  
 322 lobe system (Fa1g). The letters refer to remarkable architectural elements that characterize the Homewood system: a detailed view and explanation of each of them is available in Figure 12. The colors used in the stereoplots correspond to the location circles presented in the outcrop model  
 323 (top). Stereoplots (Schmidt, lower hemisphere) highlight the paleocurrents taken in the turbidite system (Fa1g, in green) and the fold axis of the slump-related folds taken in the mass-wasting system (Fa3p, in purple) after back-tilting of bedding planes to initial horizontal position (assuming  
 324 cylindrical folding). The blue numbers U1 to U6 highlights six units, corresponding to six distinct episodes of mass wasting. (B): Sedimentary section 01 (SS-01) recorded at Homewood. This section provides an overview of the three facies associations encountered.







326 Figure 10: [previous page] Detailed views of the shelf-derived gravity-driven deposit main architectural and sedimentary  
327 elements supporting the interpretation of the Waikaraka 3D outcrop model (Figure 8). Fa2c corresponds to the confined slope  
328 channel system facies association, whereby Fa2c-e means early-stage fills and Fa2c-l late-stage fills. CF, DF, gHDTC,  
329 gsHDTC, SST, HDTC, LDTC correspond to the lithofacies summarized in Table 1. (a): Discontinuity *D5* from [Bailleul et al.](#)  
330 [\(2007\)](#) and [Bailleul et al. \(2013\)](#) marking the transition from Fa2c-e to the uppermost sand-rich deposits; (b, e): Fa2c-l fining-  
331 and thinning-upward conglomerate (gHDTC, gsHDTC) to sandstone series (HDTC, LDTC) overlain by Fa2c-e amalgamated  
332 conglomerates (CF and gHDTC) then followed by sandstones (HDTC, LDTC); (c): Fa2c-l boulder-grade clast imbrication  
333 (base of gsHDTC); (d): Fa2c-l three successive channel storey fills characterized by conglomerate-rich bases (gsHDTC) with  
334 upward (*e.g.*, flame) and downward (*e.g.*, load) soft-sediment features, overlain by sandstone-rich lithofacies (HDTC to  
335 LDTC); (f, h): Fa2c-l amalgamated, broadly lenticular channel storey fills (gsHDTC, HDTC, LDTC) ; (g): Fa2c-l upward (*e.g.*,  
336 flame) soft-sediment features and trough cross stratifications; (i): Fa2c-e transitioning to Fa2c-l; (j): Fa2c-e amalgamated  
337 sandstones (HDTC) pinching out to the north overlaid by the Fa2c-e conglomerates (DF, CF); (k): large-scale fluid escape  
338 structures and downward (*e.g.*, load) soft-sediment features in the Fa2c-e sandstones series (HDTC, LDTC) that underlies  
339 the next Fa2c-e conglomerate events (DF, CF); (l): Fa2c-e amalgamated conglomerates with outsized clasts (CF).

340 The deposits are moderately-sorted and in average comprise sub-rounded granule- to pebble-grade  
341 clasts with a few scattered boulders and outsized clasts (Figure 4; see Section 6). To the north, these  
342 conglomerates are matrix-supported (Figure 8) and locally include a very thin, clast-supported basal layer  
343 (**sheared DF**) (Figure 4). The overlying sandstones characteristically present well-developed, southward-  
344 oriented soft-sediment deformation features in their structured intervals and their beds are frequently  
345 contorted, presenting apparent southward-verging recumbent folds (Figure 8). To the south, these folds  
346 appear to eventually evolve into matrix-supported conglomerates containing both folded sandstones and  
347 extraformational pebble-grade clasts (**SL-a to DF**) (Figure 8; Figure 11).

348 Finally, the third occurrence is present toward the top of the Waikaraka outcrop and alternatively  
349 comprises disorganized (**CF**) and organized (**gHDTC**) conglomerates, sometimes overlain by structured  
350 to structureless sandstones (**SST**) (Figure 8; Figure 10e; Figure 11). The deposits are generally  
351 moderately sorted and, on average, comprise sub-rounded pebble- to cobble-grade clasts (Figure 4; see  
352 Section 6), although some occurrences carry boulders and outsized clasts (Figure 10e). The overlying  
353 sandstone series is only a few meters thick, interrupted by another couple of meter-thick conglomeratic  
354 episode (fourth occurrence), itself characterized by a sharp upper surface (Figure 10a).

### 355 Interpretations

356 **Fa2c-e** is interpreted to characterize the axis to off-axis, early-stage (Stage I of [Kneller et al. \(2020\)](#)) fills  
357 of confined slope channel complexes (*sensu* [Sprague et al. 2005](#)).

358 The laterally extensive, coarse-grained and clast-supported conglomerates at its base represent  
359 deposition from gravelly gravity flows (Lowe 1982, R1, R2, R3 divisions). Upslope failures may have  
360 triggered all or part of these flows, which then began to transform from cohesive to cohesionless debris  
361 flows downslope (Lowe 1982; Postma 1986; Sohn et al. 2002). Their abundant amalgamation surfaces,  
362 clast-supported textures (e.g., disorganized, inverse, normal) and very coarse grain sizes indicate  
363 significant bed load transport and rapid deposition of the gravels underneath gravelly high-density  
364 turbidity currents in an environment dominated by high-energy flows, erosion and bypass (Lowe 1982).  
365 The abundant scoured bases, combined with the other textural and sedimentary features suggest that  
366 these flows were funneled through low sinuosity, braid-like channels, as recurrently found in base of  
367 confined slope channel complexes (e.g., Galloway 1998; Di Celma et al. 2010; Gamberi et al. 2013; Li et  
368 al. 2018; McArthur and McCaffrey 2019; Kneller et al. 2020). The southward increase in clast size  
369 observed in the first occurrence suggests a southward evolution from off-axis to axis settings (Campion  
370 et al. 2000).

371 The general upward internal organization observed in the conglomerates and the overlying sandstone  
372 series likely resulted from gravelly to sandy high- to low-density turbidity currents from multiple, rather  
373 than single flows (Lowe 1982), with flows initiating from a shallow marine environment connected to a  
374 vegetated hinterland sourcing the abundant terrigenous material (Kuenen 1964). The resulting deposits  
375 suggest lower energy environments leading to both the progressive abandonment of the channel system  
376 and passive infilling of the relief (Galloway 1998; McHargue et al. 2011). Axis to off-axis abandonment  
377 facies are typically sand-rich, fairly thick (e.g., medium-bedded) and highly to moderately amalgamated  
378 (Campion et al. 2000; McHargue et al. 2011; Hubbard et al. 2014).

379 As previously described, four distinct occurrences of these axis to off-axis, early-stage fills were observed  
380 at Waikaraka, thereby indicating the presence of several slope channel complexes infilling one larger  
381 channel conduit (*sensu* Sprague et al. 2005). However, **Fa2c-e** only provides information on the deposits  
382 that form the lower part of these channel complexes.

383 The textural changes observed in the second occurrence here suggest transformation from cohesive to  
384 cohesionless flows, resulting from progressive dilution of a parental debris flow (DF) evolving into a high-  
385 concentration gravelly dispersion overlain by a lower-concentrated and turbulent suspension (gHDTc)  
386 (Sohn et al. 2002, *i.e.*, bipartite or stratified flows). When taken in consideration with the recurrent soft-  
387 sediment deformations, these elements suggest that this second occurrence may originate from several,  
388 small-scale failure events, either at or near the shelf break, or directly from the channel walls (Di Celma  
389 et al. 2010; Janocko et al. 2013; Janocko and Basilici 2021). The resulting mass-wasting products

390 deposited in the channel conduit and the southward component observed in the above features indicate  
391 a potential northward location of the source region. Whether it is regional or local is here a matter of  
392 speculation. The large-scale fluid escape structures present in the underlying sandstones would result  
393 from sudden loading of the mass-wasting products onto the sandstone strata, increasing pore pressure  
394 and promoting dewatering (Figure 10k) (Browne et al. 2020).

395 Despite the changes observed in the second occurrence's texture and clasts (see Section 6.3), the  
396 absence of shutdown mud drape, the only weakly erosional basal surface separating the first and second  
397 occurrences, and their general crude fining- and thinning-upward trend all suggest that these two  
398 occurrences belonged to the same channel complex, rather than two distinct complexes (Kneller et al.  
399 2020). The same reasoning applied to the third and fourth occurrences allows considering that these two  
400 belonged to another confined channel complex.

401 Overall, the lateral and vertical facies evolution observed throughout **Fa2c-e** suggests a general evolution  
402 from off-axis to axis settings towards the south. Notwithstanding the overall outcrop conditions (no 3D  
403 control) and or the orientation of the outcrop to withhold their exposures, the absence of recognizable  
404 lateral-accretion packages (LAPs) (*sensu* Abreu et al. 2003) may also inform that the slope channel  
405 system (**Fa2c**) was likely low to moderately sinuous and generally erosional. Such channel belts usually  
406 occur in the upper- to middle-parts of the continental slope (Janocko et al. 2013). Finally, the  
407 micropaleontological analyses suggest that deposition took place at outer bathyal paleodepths (Appendix  
408 2).

### 409 **5.2.2. Confined slope channel: axis to off-axis, late-stage fills (Fa2c-l)**

#### 410 Observations

411 **Fa2c-l** presents contrasting sedimentary facies and architectural styles to those of the underlying **Fa2c-**  
412 **e**, characteristically comprising fining- and thinning-upward series, dominated by fine-grained facies, only  
413 punctuated by coarser intervals. This association is traceable across the entire outcrop laterally (>900  
414 meters) and can hold over 40 meters in thickness.

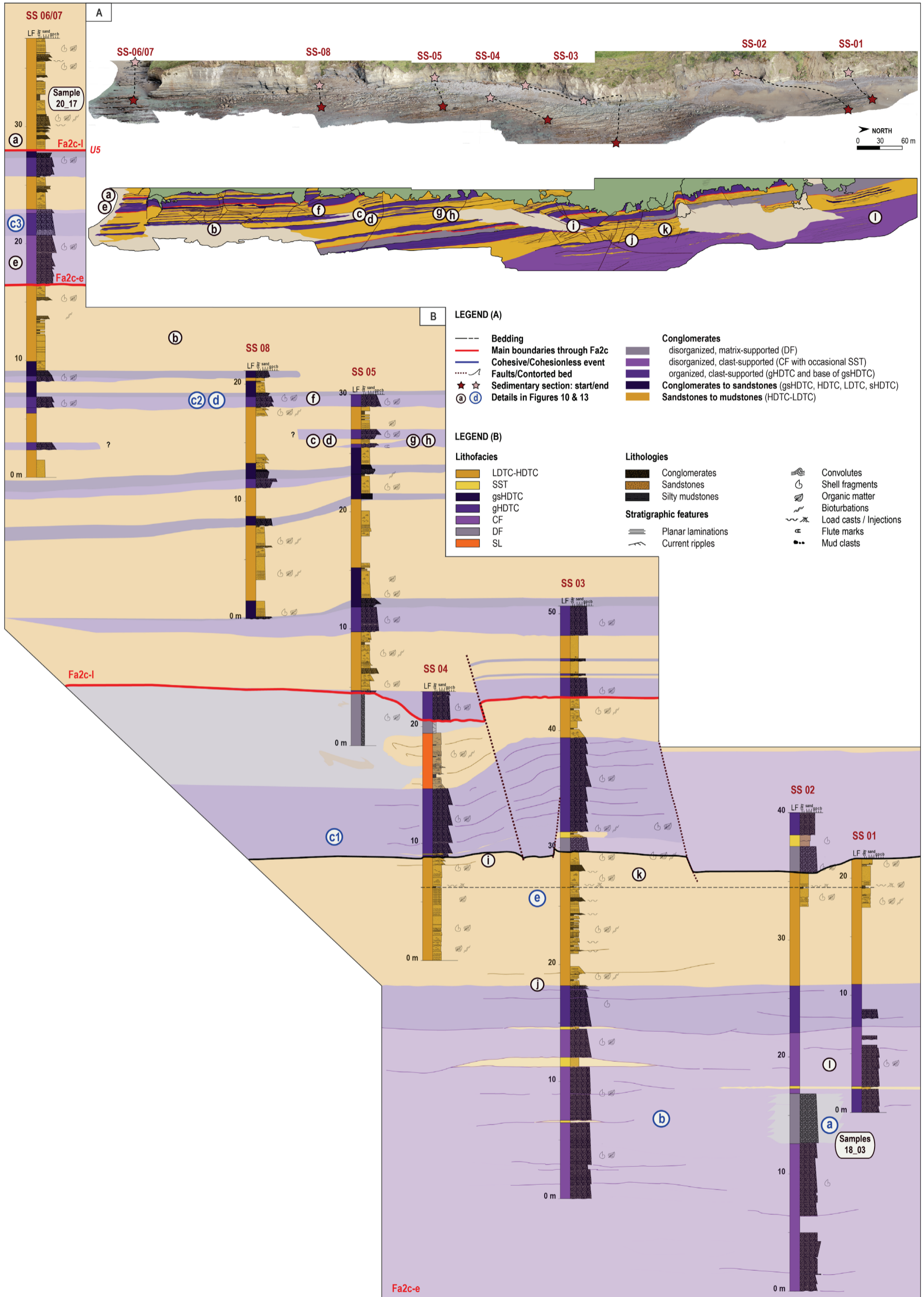


Figure 11: Sedimentary sections recorded at the Waikaraka outcrop. This system comprises two distinct channel complexes, respectively divided into the early-stage confined channel fills (Fa2c-e) and late-stage confined channel fills (Fa2c-l). The letters in dark red refer to remarkable architectural elements that characterize the Waikaraka system: a detailed view and explanation of each of them is available in Figure 10. The letters and numbers in blue give the location of the clast analyses summarized in Figure 13. Details on the sample data are available in Appendix 2. The red surface named *D5* at the top of the Waikaraka fill corresponds to the discontinuity *D5* from (Bailleul et al. (2007) and Bailleul et al. (2013), which marks the transition to the period of generalized subsidence that affected the Hikurangi Margin during the Middle Miocene, Waiuan.



1 **Fa2c-I** comprises eight successive channel storey fills (*sensu* Sprague et al. 2005) (Figure 8; Figure 11).  
2 Each storey commonly starts with amalgamated, thick- to very thick-bedded (one to one and a half  
3 meters), organized, polymict and clast-supported conglomerates (**gHDTC**) deposited above a broadly  
4 erosional, concave-up base (Figure 4; Figure 11). These are overlain by a fining- and thinning-upward  
5 conglomerate to sandstone series, capped by bioturbated mudstones (**gsHDTC**), which is, in turn,  
6 overlain by another fining- and thinning-upward series of nearly-flat bottomed, thick- to thin-bedded  
7 sandstones and mudstones (**HDTC** to **LDTC**) (Figure 5; Figure 10b, h; Figure 11). The basal  
8 conglomerate to sandstone series shows repeated bed amalgamation, thus forming composite  
9 stratigraphic infills (Figure 10f; Figure 11). Conglomerates are concentrated within the basal intervals;  
10 sandstones quickly become dominant, generally representing a third of the channel storey fill (Figure 8;  
11 Figure 11). The conglomeratic intervals are broadly lenticular, laterally pinching out, amalgamated and  
12 often include clast imbrications and subtle cross-stratifications (Figure 10c, f, h). Interestingly, their matrix  
13 seems primarily made of bioclast content (*e.g.*, shell fragments) toward the top of **Fa2c-I**.

14 The bases of the conglomerates commonly display centi- to decametric soft-sediment deformation  
15 structures (*e.g.*, flames) and large sole marks (*e.g.*, flutes) (Figure 5; Figure 10d, g). Boulders and  
16 oversized clasts usually mark the transition from the inverse- to normally-graded patterns, and their  
17 proportion generally increases toward the south (Figure 5, gsHDTC to HDTC). Clasts are sometimes of  
18 larger size than the conglomerate bed itself, thus protruding into the overlying sandstones (Figure 5,  
19 gsHDTC to HDTC). On average, sub-rounded pebble-grade clasts dominate. In **Fa2c-I**, the **HDTC**  
20 sandstones that directly overlie the conglomerates are commonly coarse-grained in their traction  
21 intervals, show trough cross-stratifications and include plant debris, numerous shell fragments (Figure  
22 10g) and abundant aligned mud clasts at their base. These sandstones form broadly lenticular bodies,  
23 with locally erosive bases characterized by very coarse- to coarse-grained material (Figure 10a, f). They  
24 eventually reach medium- to fine-grain sizes in the remaining intervals and in the subsequent fining- and  
25 thinning-upward series (**HDTC** to **LDTC**). The paleocurrents measured in the sandstone intervals indicate  
26 a general southward trend (Figure 8), which is consistent with measurements from imbrications within  
27 the coarse-grained basal fills.

28 Two occurrences of **Fa2c-I** were observed at Waikaraka (Figure 8; Figure 11).

29 The first occurrence corresponds to the above description. It is about 40 meters thick and comprises eight  
30 channel storeys (*sensu* Sprague et al. 2005). Its basal surface is sharp, concave-up and broadly incises  
31 **Fa2c-e** with up to 2 meters relief (Figure 8). The second occurrence is observed at the top of the  
32 Waikaraka outcrop and only crops out for about 15 meters laterally and vertically, being covered by

33 quaternary beach deposits after that. Its basal surface is sharp above the underlying **Fa2c-e** occurrence  
34 and it is only comprised of sandstones (**HDTC to LDTC**) (Figure 10a, e; Figure 11).

### 35 Interpretations

36 **Fa2c-l** is interpreted to represent the axis to off-axis, late-stage (Stage II of Kneller et al. (2020)) fills of  
37 confined slope channel complexes (*sensu* Sprague et al. 2005). Characterized by multiple episodes of  
38 erosional bypass and sedimentation, these form the middle fills of channel complexes.

39 The conglomerate- to sandstone-dominated basal successions (**gHDTC** and **gsHDTC**) that broadly fill  
40 the erosion surface result from stratified gravelly gravity flows that are confined within channels (Lowe  
41 1982; Campion et al. 2000; Sohn et al. 2002). The basal dewatering features suggest rapid deposition  
42 and loading of the channel fills, whereas the frequent amalgamation observed in both the conglomerates  
43 and sandstones implies that erosion and bypass were dominant in the basal interval (Lowe 1982).  
44 Transport of large (*e.g.*, oversized) clasts can be explained by flow confinements such as the ones found  
45 in confined slope channels, where the flow size, velocity and carrying capacity are maximized (Postma  
46 et al. 1988). The high degree of preservation of mud clasts in both the conglomerates and overlying  
47 sandstones, scattered as lag material indicates high fall-out rates (Postma 1986), whereas the increase  
48 in content of coarse and fine skeletal fraction toward the top of the series rather suggest deposition of  
49 calciturbidites (*e.g.*, Haak and Schlager 1989; Reijmer et al. 2015). The gravelly gravity flows then  
50 decelerated and became less confined. The overlying tabular, fining- and thinning-upward series of  
51 sandstones and mudstones (**HDTC to LDTC**) mark the transition to lower energy depositional  
52 environments, resulting in passive filling and progressive abandonment of the channel system (Galloway  
53 1998; McHargue et al. 2011). As in **Fa2c-e**, such sand-rich abandonment facies here suggest axis to off-  
54 axis settings (Campion et al. 2000). Overall, these broadly lenticular and laterally extensive channel  
55 storey fills are dominated by sandy high- to low-density turbidites, scouring a mud-rich substrate during  
56 downslope transport and or at hydraulic jumps (*i.e.*, abundant mud clasts) (Fonnesu et al. 2016). The  
57 vertically aggrading, semi-amalgamated style of the channel storey fills confirms a generally confined  
58 slope channel complex settings (Sprague et al. 2005), but less steep and confined to those of the  
59 previously described **Fa2c-e**.

60 Although most likely belonging to the same slope channel system, the uppermost occurrence  
61 characteristics and micropaleontological analyses both suggest contrasting depositional settings. Not  
62 only the micropaleontological study indicates lower paleodepths of deposition (*i.e.*, lower bathyal), it also  
63 indicates that this latter series was restricted to the Middle Miocene, Waiauian (Serravallian, 13.05 – 11.63  
64 Ma) (Figure 11; Appendix 2).

65 Overall, we interpret the Waikaraka outcrop to represent the axis to off-axis stratigraphic infills of two  
66 distinct channel complexes, essentially capturing the early- to late-stage fills of a larger, yet confined  
67 slope channel system (*sensu* Sprague et al. 2005). We further discuss the controls on such a system and  
68 the implications of its uppermost deposits in Section 7.

### 69 **5.3. Weakly confined slope and depositional lobe systems**

70 The Homewood section crops out one kilometer north of the Kaiwhata river mouth and extends over 110  
71 meters along the coast (Figure 3). It features three distinct Middle Miocene, Lillburnian (Late Langhian to  
72 Serravallian, 15.1 – 13.05 Ma, Appendix 2) gravity-driven systems described and interpreted below as  
73 three separate facies associations (**Fa2c**, **Fa3p**, **Fa1g**) (Figure 9).

#### 74 **5.3.1. Slope channel (Fa2c)**

##### 75 Observations

76 At Homewood, **Fa2c** is characterized by crude fining- and thinning-upward series, evolving from coarse-  
77 to fine-grained facies. Owing to outcrop conditions, it is intermittently exposed for about 80 meters  
78 laterally and only represents a ten of meter-thick succession of sub-vertical beds. Its basal surface does  
79 not crop out and it is overlain by the erosive **Fa3p** (Figure 9).

80 Here, **Fa2c** is characterized by a series of tabular, continuous (1) medium- to thick-bedded (10 to 50  
81 centimeters), disorganized (**CF**) to organized (**gHDTC**) clast-supported conglomerates with sandstone  
82 caps (**SST**), followed by (2) thin- to medium-bedded (3 to 15 centimeters) pebbly sandstones (**sHDTC-a**)  
83 or bioclastic grits (**sHDTC-b**), either interbedded with mudstones (**MDST**) or structureless sandstones  
84 (**SST-b**) and mudstones (**MDST**), finally overlain by (3) medium- to thick-bedded (10 to 50  
85 centimeters) sandstones with mudstone caps (**LDTC-b**) (Figure 6; Figure 7; Figure 9; Figure 12g, h).

86 Overall, a greater internal organization is observed in **Fa2c** to the north-east of the outcrop.  
87 Conglomerates are generally ungraded (**CF**) to the south-west of the outcrop, whereas they gradually  
88 show internal organization (**gHDTC**), displaying inverse- to normally- or only normally-graded patterns to  
89 the north-east (Figure 6). The conglomerates mostly consist of sub-angular to sub-rounded granule- to  
90 pebble-grade clasts (see Section 6). Cobbles and boulders are locally observed. Similarly, the bioclastic  
91 grits (**sHDTC-b**) become normally-graded, and include mud clasts and or bioclastic fractions aligned  
92 along horizons towards the north-east of the outcrop (Figure 7). In both cases, the basal surfaces of the  
93 conglomerates and bioclastic grits are slightly erosive and commonly display load structures.

94 Finally, the sandstones (**LDTC-b**) to the south-west present rare, discontinuous, centimeter-thick, very  
95 coarse basal intervals (Ta), whereas to the north-east, these basal intervals are well-developed, fairly  
96 continuous and include frequent mud clasts and bioclastic fractions aligned along horizons, as well as  
97 load structures (Figure 7). The related sandstones also commonly show distinct grain-size breaks,  
98 indicating amalgamation of beds.

### 99 Interpretations

100 We interpret this basal succession to result from both flow processes and depositional environments  
101 (**Fa2c**) similar to those described at Waikaraka. While outcrop exposure hinders recognition of specific  
102 architectural elements, such as channels, and of particular depositional units (**Fa2c-e**, **Fa2c-l**), the  
103 stratigraphic infill outlines comparable crude fining- and thinning-upward conglomerate to sandstone  
104 series, resulting from gravelly to sandy high- to low-density turbidity currents (Lowe 1982). The main  
105 difference resides in the scale of the individual beds, being significantly smaller at Homewood, which  
106 could either point towards (1) smaller source region or slope-channel system, and or (2) contrasting  
107 locations along the slope or laterally within the channel.

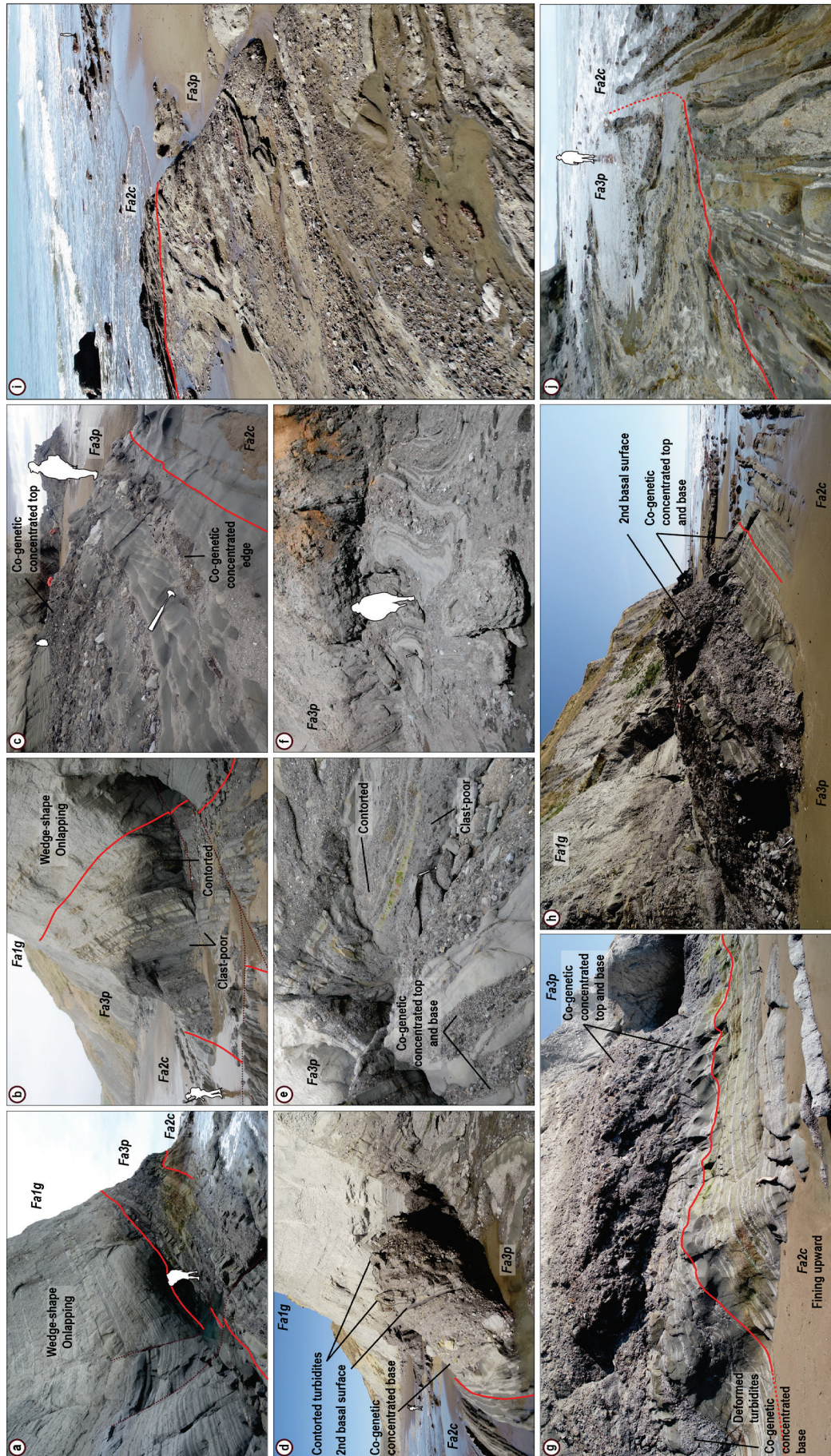
### 108 **5.3.2. Mixed shelf-derived mass-transport deposits (Fa3p)**

#### 109 Observations

110 **Fa3p** mostly comprises interbedded coarse-grained facies, here exposed across more than 120 meters  
111 laterally and about 15 meters thick (Figure 9).

112 **Fa3p** is characterized by a succession of thick- to very thick-bedded (<one to >two meters), disorganized,  
113 polymict, matrix- (**DF**) and clast-supported (**CF**) conglomerates, that laterally becomes entirely deformed  
114 and contorted (**SL-a**) to the north-east of the outcrop (Table 1; Figure 6; Figure 7; Figure 9). The  
115 conglomerates are recurrently capped by thin- to thick-bedded, either structureless or structured  
116 sandstones (**SST**) (Table 1; Figure 6; Figure 7; Figure 9; Figure 12), the latter of which can sometimes  
117 evolve to a few interbedded structured, very thin- to medium-bedded sandstones with mudstone caps,  
118 commonly bioturbated (**LDTC-a**) (Table 1; Figure 7; Figure 9; Figure 12b, e).







120 Figure 12: [previous page] Detailed views of the shelf-derived gravity-driven deposit main architectural and sedimentary  
121 elements supporting the interpretation of the Homewood 3D outcrop model (Figure 9). Fa2c corresponds to the confined  
122 slope channel system facies association, Fa3p to the shelf-derived mass-wasting system and Fa1g to the distal depositional  
123 lobe system. SL, CF, DF, gHDTc, SST, sHDTc, HDTc, LDTc correspond to the lithofacies summarized in Table 1. (a, b):  
124 Fa3p succession of interbedded conglomerates (DF, CF) recurrently capped by sandstones (SST) and overlain by the  
125 onlapping Fa1g sandstones (HDTc, LDTc); (b): Fa3p varying bed dips and upper contorted unit; (c, d, e, g, h): Fa3p  
126 conglomerates showing local variation in clast concentrations at their base, edge and or top (sheared DF), (d, h): Fa3p sharp  
127 surface at the base of the succeeding conglomerate, which holds contorted sandstones; (e): Fa3p occurrence of clast-poor  
128 conglomerate (DF) and upper contorted unit; (f, i, j): Fa3p contorted conglomerates (SL-a, made of DF, CF and SST); (g, h):  
129 Fa2c crude fining- and thinning-upward conglomerate to sandstones (gHDTc, sHDTc, LDTc), overlain by Fa3p, which is  
130 characterized by erosion as well as local substrate entrainment and deformation at its base; (i, j): Fa3p basal eroded through  
131 Fa2c towards the north-east.

132 The basal surface is highly erosive (>8 meters) towards the north-east (Figure 9, red surface at the base  
133 of Fa3p; Figure 12g, i, j), and both local entrainment and deformation (*i.e.*, apparent northward-verging  
134 contorted sandstones) of the substrate are observed (Figure 9; Figure 12g). In contrast, the upper surface  
135 is highly irregular, with overlying beds onlapping on a deci- to decametric scale (Figure 9; Figure 12a, b).  
136 Laterally, the upper surface can appear rather flat and overlain by relatively tabular sandstones. The first  
137 overlying sandstone beds are generally internally chaotic; they sometimes incorporate material from **Fa3p**  
138 (*e.g.*, pebbles and cobbles) and can form well-developed, decametric-scale soft-sedimentation  
139 deformation structures.

140 Five additional internal surfaces were also identified, which allows division of **Fa3p** into six units (Figure  
141 9, blue surfaces within Fa3p). Commonly observed at the top of the structureless or structured beds  
142 (**SST**), these surfaces are sharp and characterized by thin- to medium-bedded silty mudstones (**MDST**)  
143 that can be traced across the outcrop (Figure 9, blue surfaces; Figure 12d, h).

144 In the southern and central parts of the outcrop, **Fa3p** comprises fairly tabular beds. Their thicknesses  
145 tend to increase north-eastward, thus generally forming lens-like shapes opening towards the north-east  
146 (Figure 9). The beds' orientation changes from N055 to N035 and the dip measurements show an abrupt  
147 evolution from 90° (sub-vertical) to ~60° towards the west in only 15 meters. The dip of the first four units  
148 (separated by the internal surfaces described above) changes from 90° to 78°, whereas the remaining  
149 two directly onlap onto the preceding ones with a dip starting around 70° and evolving towards 60°. To  
150 the north-east, **Fa3p** is completely disorganized and shows SW-NE oriented recumbent folds (Figure 9,  
151 slump measurements; Figure 12j).

152 Three different scales of internal contortion are observed in **Fa3p**, evolving both vertically and laterally:  
153 (1) at clast-scale; (2) at unit-scale and (3) at the scale of several units. First, the entrained sandstones

154 display both apparent NE-verging (Figure 9) and NE-SW-oriented recumbent folds (Figure 12f). Second,  
155 the fifth unit, previously observed to onlap on the underlying beds, appears to be severely folded, with a  
156 NE-SW main axis (Figure 9, unit 5; Figure 12b, e). Third, the upper **Fa3p** units become remobilized  
157 laterally, folding following an apparent NE-verging trend (Figure 12f). To the north-east of the outcrop,  
158 this remobilization involves the entire facies association (*i.e.*, interbedded **DF**, **CF** and **SST**) (Table 1;  
159 Figure 9; Figure 6; Figure 12f, i, j).

160 The matrix of the interbedded polymict conglomerates generally represents 40 to 70% of the **DF** deposits  
161 with two occurrences comprising up to 80% of matrix (Figure 6; Figure 12b, e; Figure 13), and <5 to 15%  
162 in **CF**. **DF** can display slight upward increase in matrix content. Alternatively, **DF** commonly shows vertical  
163 and or lateral variations in both clast concentration and size at its base, top and or edges that appear  
164 akin to the clast-supported lithofacies (**CF**) (Figure 9; Figure 6; Figure 12c, d, e, g, h). These concentrated  
165 intervals are characterized by clasts of smaller grades (*i.e.*, granule- to pebble-grade) than those of the  
166 general matrix-supported intervals (*i.e.*, pebble- to cobble-grade) (Figure 6). Overall, the average clast  
167 size is of pebble- to cobble-grade in **DF**, whereas granule- to pebble-grade dominates **CF**. In **DF**, larger  
168 clasts are frequent and can be outsized (deci- to decametric); conversely, they are rare and of cobble-  
169 grade in **CF**. In both **DF** and **CF**, clast edges typically vary between sub-angular to sub-rounded shapes.  
170 The highest proportions of sub-angular clasts are found in the largest clasts (*e.g.*, boulders). Sub-rounded  
171 to rounded clasts are mostly present in the concentrated granule- to pebble-grade intervals of **DF** or in  
172 some rare occurrences of **CF**.

### 173 Interpretations

174 We interpret **Fa3p** to represent deposition (*i.e.*, mass-transport deposits [MTDs]) from successive mass-  
175 wasting events that triggered recurrent debris flows (**DF** and **CF**) (Postma 1986; Mulder and Cochonat  
176 1996). The clast-supported debris flows (**CF**) are inferred to result from cohesionless flows (Postma  
177 1986). Even if matrix strength dominates in **DF**, thereby suggesting cohesive flows (Nardin et al. 1979;  
178 Lowe 1982), Nemeč and Steel (1984) demonstrated that the predominance of matrix in conglomerates  
179 was not always a criterion to discriminate cohesive and cohesionless debris flows. Although the  
180 ungraded, disorganized and matrix-rich occurrences of **DF** result from cohesive flows (Johnson 1984;  
181 Postma 1986), the 2-to-3 phase occurrences (**sheared DF**) show internal organization that indicates fully  
182 sheared flows with additional support mechanisms, such as grain collisions and dispersive pressure,  
183 characteristic of cohesionless debris flows (Lowe 1982; Nemeč and Steel 1984; Postma 1986).

184 Multiphase debris flows require steep slopes (Postma 1986). The lens-like shapes observed in some of  
185 their deposits may suggest channelization, characteristic of steep slope settings (Ortiz-Karpf et al. 2017;

186 Calhoun and Clague 2018). The high-amplitude variations recorded in the dip of beds (~40°) indicates  
187 that they were syn-kinematic with the rise of a structure to the east, *i.e.*, growth strata. Changes in slope  
188 gradient chiefly impact flow behavior (*e.g.*, velocity) and may promote flow transformations (*e.g.*, dilution,  
189 shearing) (Fisher 1983), which in turn control the nature and geometry of the resultant deposits. Hence,  
190 slope angle changes could partly explain the variety of debris flow deposits recorded in **Fa3p**. Also, the  
191 considerable amounts of granule- and pebble-grade clasts (Sohn et al. 2002) as well as the recurrent  
192 presence of mudstone layers and fine sandstones (**SST**), respectively under- and overlying the  
193 conglomerates, all point toward fast-moving debris flows that hydroplane (*sensu* Mohrig et al., 1998). The  
194 multiphase debris flows described here would thus result from efficient dilution of cohesive, hydroplaning  
195 debris flows (Mohrig et al., 1998); dilution which can arise only a few kilometers from the source areas  
196 (Sohn et al. 2002) and which can lead to the generation of slower-moving suspended sediment clouds  
197 eventually depositing above them (**SST**) (Mohrig et al. 1998; Sohn et al. 2002; Mohrig and Marr 2003;  
198 Talling et al. 2004). The first debris flow, however strongly erosional, suggests both an abrupt change in  
199 the depositional settings to that of the underlying **Fa2c** deposits and an initially different mechanism (*i.e.*,  
200 substrate erosion) responsible for flow mobility (Sobiesiak et al. 2018).

201 The internal surfaces allow to distinguish at least six mass-wasting events. The different scales of internal  
202 contortion and deformation observed in the **Fa3p** deposits however complicates their depositional history,  
203 thereby questioning the hypothesis of simple deposition of one mass flow on top of another. Basal  
204 deformation of the substrate linked to the passage of an overriding flow is common, however the flow  
205 mobility mechanisms proposed here (*i.e.*, hydroplaning) diverges from such a scenario, promoting little  
206 to no interaction between the moving flow and underlying deposits for five of the units (Sobiesiak et al.  
207 2018). Notwithstanding the possibility for the contorted strata to shear and deform in the direction of the  
208 flow and thus capture the paleoslope direction at the clast-scale (Fonnesu et al. 2016), we here suggest  
209 a different mechanism to explain the deformation recorded at the unit-scale and propose that a change  
210 in slope gradient can both trigger development of a mass flow upslope and coeval remobilization  
211 (translation and deformation) of the previously deposited and only poorly lithified mass-flow deposits  
212 downslope. When combined with bed dips, the analysis of the fold hinges and axial planes at both clast-  
213 and unit-scales consistently indicates a NW-dipping paleoslope and transport direction (Alsop et al. 2019;  
214 Alsop and Weinberger 2020), coherent with a rising structure to the east.

215 The larger-scale remobilization observed to the north-east that involves the entire facies association **Fa3p**  
216 could also be associated with this growing structure. As such, we interpret it to result from local failure of  
217 the entire series due to the recurrent loading of sediments onto this rising slope, two mechanisms known



218 to greatly affect slope stability and encourage its failure (Hein and Gorsline 1981). Finally, the  
219 micropaleontological studies imply an upper- to mid-bathyal depositional environment (Appendix 2).

### 220 5.3.3. Distal depositional lobe system (Fa1g)

#### 221 Observations

222 **Fa1g** is characterized by laterally continuous, on average medium-bedded (~14 centimeters) and fine-  
223 grained sandstones that show decreasing bed dips upward (56° to 32°). This association is exposed  
224 along ~one kilometer of coastline and is over 200 meters in thickness (Appendix 4). Only the lowermost  
225 part of this system is described in context of the outcrop model presented in this study (Figure 9; Fa1g).

226 **Fa1g** mostly comprises interbedded tabular, medium- to thick-bedded, frequently amalgamated  
227 sandstones (**LDTC-b**) and very thin- to medium-bedded (**LDTC-a**), often dewatered, argillaceous  
228 sandstones (Table 1; Figure 7). Most beds include shell fragments and plant debris in the structured  
229 intervals, and are capped by silty mudstones that regularly include bioturbations (Figure 7) such as  
230 *Phycosiphon* sp., *Zoophycos* sp., *Skolithos* sp. and *Scolicia* sp. Episodes of thin-bedded bioclastic grits  
231 (**sHDTC-b**) are recurrent (Figure 7). Very-thick sandstone beds punctually occur; they tend to have sole  
232 marks, several grain-size breaks, and sometimes show decametric soft-sedimentation deformation  
233 structures. A few lenticular, medium- to very thick-bedded sandstones (**HDTC**) are also observed. Their  
234 incision rarely exceeds 50 centimeters and can comprise coarse bioclastic material (Figure 7).  
235 Paleocurrent measurements from sole marks and traction structures generally indicate a north-eastward  
236 direction at the base of this system (Figure 9). Finally, disorganized, contorted strata (**SL-a**) and coherent,  
237 remobilized strata displaying sharp internal truncations in the same lithofacies (**SL-b**) recurrently interrupt  
238 the strata (Table 1; Figure 6).

#### 239 Interpretations

240 **Fa1g** is interpreted to record the emplacement and sustained development of a distal depositional lobe  
241 system (Bailleul et al. 2007; Pr elat et al. 2009; McArthur et al. 2021) installed above and subsequently  
242 pinned at its base within the rugose topography left by the underlying MTDs (e.g., Bull et al. 2009;  
243 Armitage et al. 2009) (Figure 9; Figure 12a, b). These laterally extensive, tabular, sandstone-rich deposits  
244 are interpreted as the deposits of dominantly low-density turbidity currents (**LDTC**) (Bouma 1962),  
245 commonly deposited in off-axis to fringe positions (Pr elat et al. 2009). The punctual bioclastic grits as  
246 well as the locally erosive, amalgamated and very thick occurrences, are however attributed to higher-  
247 density turbidity currents (Lowe 1982) resulting from larger flow events. The onlapping and growth strata

248 geometries as well as the recurrent yet localized remobilization of strata both suggest an unstable  
249 depositional environment, likely due to a continued rise of the eastern structure. The paleocurrent  
250 directions to the NE indicate that the flow at the start of this depositional system was parallel to the rising  
251 NE-SW oriented structure to the east and thus both point towards a different sourcing system to the  
252 underlying MTDs (see Section 5.3.2) and an axial routing of the turbidity currents (Burgreen and Graham  
253 2014). The abundance of land-derived material suggest that these turbidity currents likely initiated from  
254 a shallow marine environment connected to a hinterland (Kuenen 1964), while the micropaleontology  
255 analyses indicate that deposition of the turbidites occurred at mid bathyal depths or deeper (> 800 meters)  
256 (Appendix 2).

## 257 6. SOURCE REGIONS

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258 Despite lithofacies similarities, the sediment gravity flows observed at Waikaraka and Homewood are  
259 lithologically different (Figure 13) and thus indicate different sediment sources. Here, we use the nature  
260 of the reworked material in the conglomerates (**DF**, **CF**, **gHDTC**, **gsHDTC**), pebbly sandstones (**sHDTC-**  
261 **a**), bioclastic grits (**sHDTC-b**) and coarser intervals of the sandstones (**HDTC**) as a tool for better  
262 understanding and characterizing their provenance and source areas as well as deciphering local  
263 paleotransport directions (e.g., Nemeč and Steel, 1984).

### 264 6.1. Extrabasinal sources

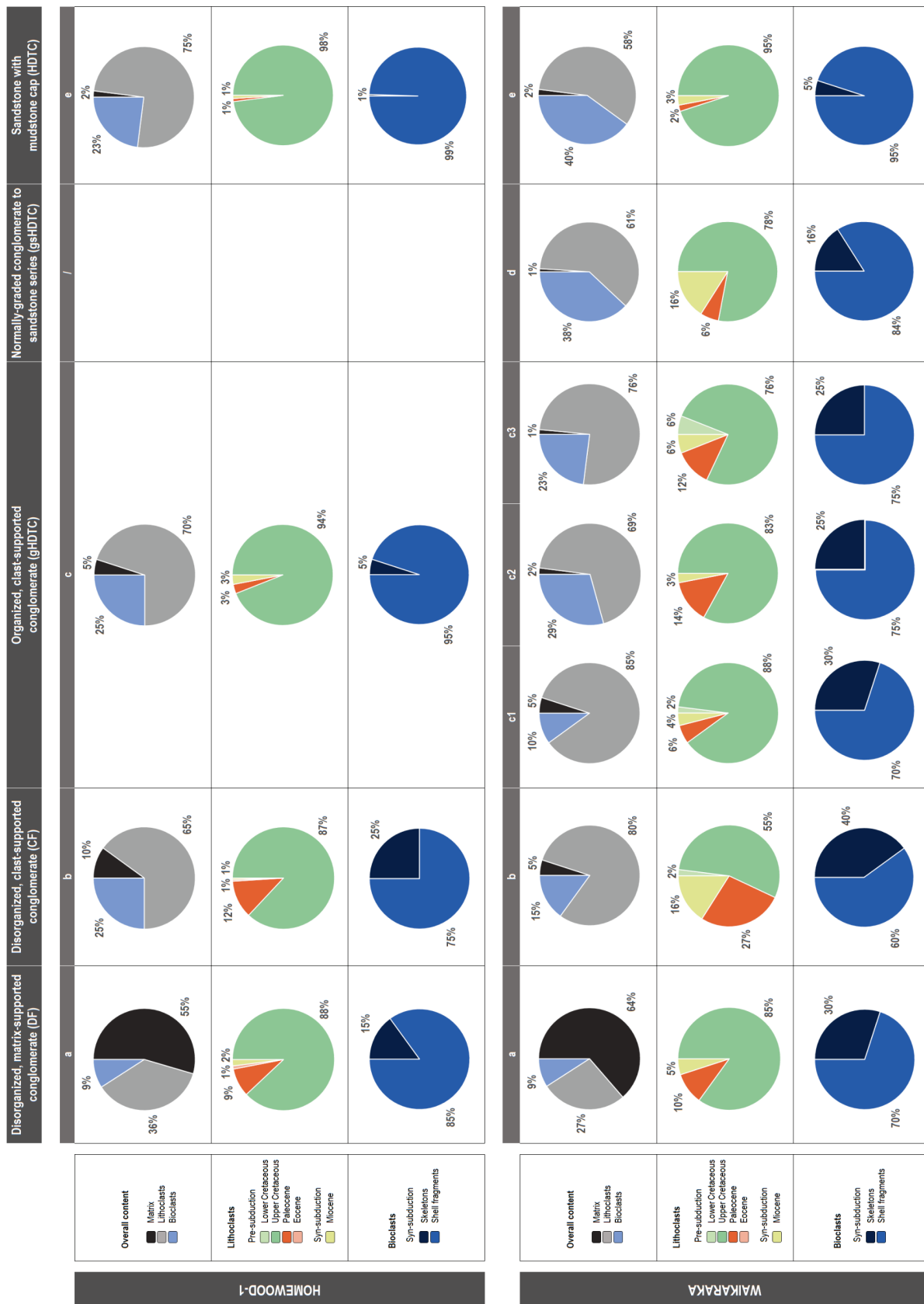
#### 265 Clast observations

266 Across both Waikaraka and Homewood, the conglomeratic units (**DF**, **CF**, **gHDTC**, **gsHDTC**, **sHDTC**,  
267 **HDTC**) essentially rework extraformational clasts from pre- and syn-subduction strata (>95%) (Figure  
268 13). Pre-subduction material dominates (> 80%) and comprises lithoclasts from the Early Cretaceous up  
269 to the Paleocene at Waikaraka and from the Late Cretaceous up to the Eocene at Homewood. The syn-  
270 subduction material is divided into Miocene lithoclasts and bioclasts.

271

272

273





274 belonged to and the bioclasts were subdivided into two main types, *i.e.*, the preserved skeletons and the shell fragments. DF,  
275 CF, gHDTC, gsHDTC and HDTC correspond to the lithofacies summarized in Table 1.

276 Intraformational clasts are virtually absent (<1%) in **CF**, but are more frequent in the conglomerates  
277 deposits resulting from matrix strength (**DF**; *e.g.*, poorly lithified turbidites) and fluid turbulence (**gHDTC**,  
278 **gsHDTC**, **sHDTC**, **HDTC**; *e.g.*, mud clasts).

### 279 Clast interpretations

280 The dominance of extraformational clasts implies that the flows responsible for these conglomeratic units  
281 initiated outside of their final depositional environment and incorporated material belonging to structural  
282 units and or paleogeographic domains different to that of the intrabasinal sediments (*i.e.*, extrabasinal)  
283 (*e.g.*, Ogata et al. 2019). The abundance of pre-subduction material indicates that during the Middle  
284 Miocene, the exposed extrabasinal paleogeography was dominated by Cretaceous to Eocene strata. In  
285 compressional settings, uplift of thrust-related structures is known to develop bathymetric highs and or  
286 steep slopes, which can in turn, expose rocks from the older substratum (*i.e.*, pre-subduction), ready to  
287 be reworked and incorporated into the subsequent flows and MTDs (*e.g.*, Ogata et al. 2019).

## 288 **6.2. Mixed siliciclastic-carbonate shelves**

### 289 Bioclast observations

290 A substantial amount (~10 to 40%) of syn-subduction bioclasts is present in all the deposits, either as  
291 shell fragments or as well- to partly-preserved skeletons (Figure 13). Molluscan species (*e.g.*, bivalves  
292 and gastropods) dominates; yet, corals (solitary and colonial) and red algae are also locally present  
293 (Appendix 3). At Homewood, the preserved forms of macrofossils reach between 15 and 25% of the  
294 bioclasts contained in the debris flow deposits (**DF** and **CF**) whereas they are virtually absent in the  
295 turbidites (**gHDTC** and **HDTC**), which are dominated by shell fragments. At Waikaraka however, the  
296 preserved forms represent a considerably larger proportion of the bioclasts, mostly oscillating between  
297 25 and 40%, even at the base of the turbidites (**gHDTC** and **HDTC**). Although the content in both the  
298 coarse (macrofossils) and fine (shell fragments) skeletal fractions gradually increases toward the top of  
299 the Waikaraka deposits, this upward evolution is particularly remarkable in the late-stage fill deposits  
300 (**Fa2c-I**) (Figure 13, cf. c2, c3, d, e).

301 Overall, the taphonomic analyses shows that destructive processes dominate the taphonomic record of  
302 the deposits, constructive taphonomic processes such as encrustation being virtually absent (Figure 14).  
303 At Homewood, the deposits are characterized by bioclasts being: (1) moderately to highly fragmented,

304 (2) moderately abraded and, (3) poorly bioeroded. At Waikaraka, the finer bioclastic fractions compare  
305 markedly in both debris flows (**DF**, **CF**) and high-density turbidite flows (**gHDTC**, **gsHDTC**). By contrast,  
306 the coarser skeletal material in the former exhibit higher degrees of abrasion and bioerosion than in the  
307 latter (Figure 14).

### 308 Bioclast interpretations

309 The nature and amount of reworked molluscan species indicate that the sediment gravity flows originated  
310 from shallow marine middle- to inner-shelf neritic environments before being transported to deep-marine  
311 outer to mid bathyal depths (Appendix 2; Appendix 3).

312 The contemporaneously developing mixed siliciclastic-carbonate environments (*i.e.*, compositional  
313 systems (*sensu* Chiarella et al. (2017))) previously identified and described in the Coastal Ranges present  
314 markedly similar faunal assemblages (Crundwell 1987; Chanier 1991; Bailleul et al. 2007; Bailleul et al.  
315 2013; Bailleul et al. 2015) and are thus great candidates for providing both the preserved skeletons and  
316 shell fragments. The depositional system envisaged for those sources is one of a regional, continent-  
317 attached shelf domain, recurrently supplied with terrigenous sediments from the exposed Cretaceous  
318 basement located to the west (Crundwell 1987; Chanier 1991; Bailleul et al. 2007; Bailleul et al. 2013;  
319 Bailleul et al. 2015; Clausmann et al. 2021).

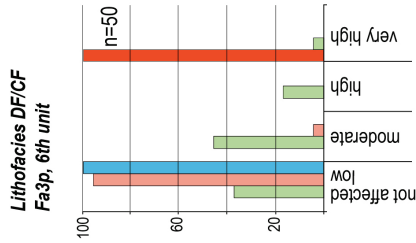
320 The substantial amount of gravels transported beyond the shelf and captured within the deposits (see  
321 Sections 5 and 6.3) requires narrow shelf settings to easily connect to the slope beyond (Nemec and  
322 Steel 1984), which typify the shelfal systems along active margins (Sømme et al. 2009). No lateral  
323 continuity has yet been proven between the mixed siliciclastic-carbonate environments and therefore,  
324 isolated, continent-detached shelfal systems likely also developed locally (*e.g.*, (Bailleul et al. 2013),  
325 Fingerpost shelf) upon and around actively growing folds and faults (Caron et al. 2004) (see Section 6.3  
326 and 7).

327 In fact, at Waikaraka, the fairly constant proportion of shelfal macrofossils found throughout the 90 meter-  
328 thick succession points to paleophysiographies allowing for molluscan communities to thrive above  
329 storm-wave base (Pomar 2001). At Homewood however, the macrofossils are mostly found in the  
330 conglomerates (**DF**, **CF**) of the mass-wasting system (**Fa3p**) and their quantity tends to be generally  
331 smaller to that of Waikaraka. This could indifferently indicate a smaller-sized source area, slightly different  
332 paleoenvironmental conditions at the source as well as different transport processes and distances. The  
333 paleotransport directions recorded in the mass-wasting system (**Fa3p**) of Homewood also inform that the

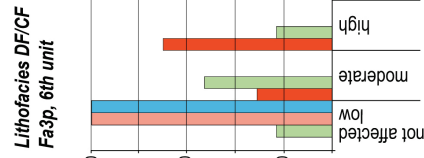
334 shelfal source was located to the east and thus was most-likely detached from the main regional shelf  
 335 domain located to the west.

**HOMEWOOD OUTCROP (Fa3p)**

**Bioclastic fraction (0.5 cm-3 cm)**

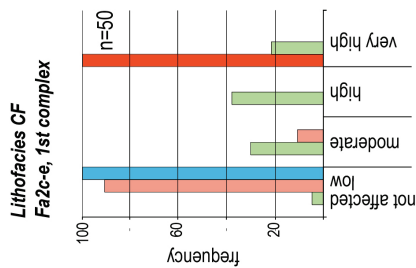


**Bioclastic fraction (>3 cm)**

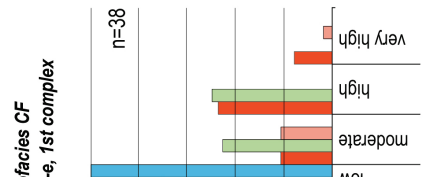
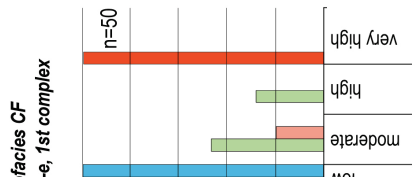
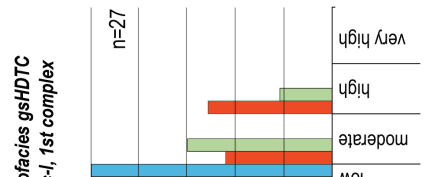
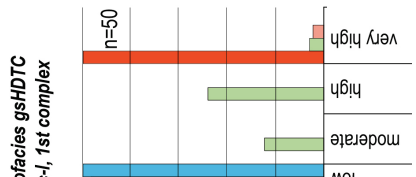
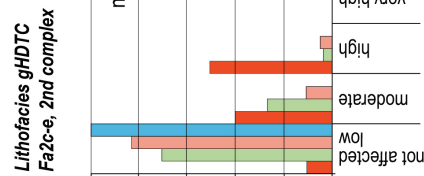
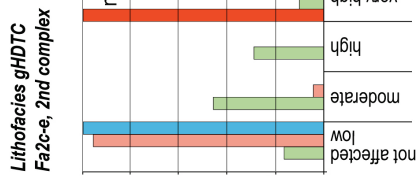
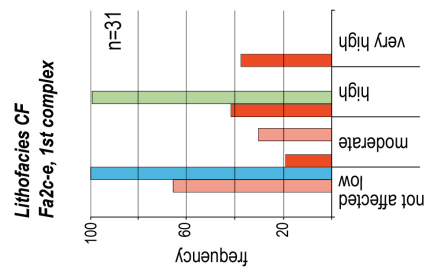


**WAIKARAKA OUTCROP (Fa2c)**

**Bioclastic fraction (0.5 cm-3 cm)**



**Bioclastic fraction (>3 cm)**



- LEGEND**
- Fragmentation
  - Abrasion
  - Bioerosion
  - Encrustation

336



337 Figure 14: Taphonomic characterization of fossil remains described from one square meter area of outcrop at Waikaraka and  
338 Homewood, using frequency histograms of the degree of alteration (e.g., low, moderate, high, very high) for each category of  
339 skeleton damage (i.e., fragmentation, abrasion, bioerosion) and encrustation. Lithofacies CF, DF, gHDTC and gsHDTC are  
340 described in details in Table 1 and facies associations Fa2c and Fa3p in Section 5.

341 Conversely, the profuse shell fragments found in the overlying, well-developed, north-eastward  
342 prograding turbidite system (**Fa1g**) indicate a change of source region, likely belonging to the main  
343 westward continent-attached shelf domain.

344 The regular detrital sand and pebble supply (see Sections 5 and 6.3) likely restricted the development of  
345 carbonate factories dominated by suspension-feeding organisms (Caron et al. 2004), thereby promoting  
346 molluscan communities (Bimol assemblage of Hayton et al. (1995)) often associated with relatively turbid  
347 waters due to fine material in suspension. The occurrence in the deposits of occasional coralline red algal  
348 branches and small-sized-rhodoliths indicate sources within the photic zone (Pomar 2001; Nebelsick et  
349 al. 2005).

350 Differences in the degrees of taphonomic alteration observed in skeletal remains relate to their residence  
351 time in the taphonomic active zone (TAZ; (Davies et al. 1989)). Typically, taphonomic alteration  
352 consistently increases with increasing residence time in the TAZ where skeletons are affected by  
353 destructive processes, such as disarticulation, fragmentation, abrasion and bioerosion. For example, low  
354 net sedimentation rates and repeated exhumation of buried skeletal remains by waves and currents, and  
355 by seafloor burrowers favor mechanical destruction and bioinfestation. By contrast, rapid and definite  
356 burial prevents such alterations (e.g., Best and Kidwell 2000; Best et al. 2007). Fragmentation of skeletal  
357 remains over three centimeters in size is moderate to high, both at Waikaraka and Homewood in debris  
358 flows and turbidite flows (Figure 14). Although it cannot be ruled out that skeletons were partly fragmented  
359 in their source area through hydraulic and biological processes (Zuschin et al. 2003), it should be  
360 expected that fragmentation also occurred, as a result of mechanical crushing, once they were integrated  
361 in their host clast-supported gravity-driven flows. Abrasion of the coarser skeletal fraction, translating in  
362 the loss of ornamentation and rounding of bioclast edges, is: (1) higher in debris flows than in turbidite  
363 flows, and (2) overall higher than fragmentation in debris flows compared to turbidite flows. Abrasion is  
364 caused by physical erosion, through friction and at a lesser degree through waterborne collision, and  
365 reworking leading to mobilization and redeposition of bioclasts. Moderate to high degrees of abrasion,  
366 along with the abundance of small-sized skeletal debris, are indicative of frequently destabilized  
367 substrates by wave action, storms and, potentially here, gravity-driven flows (e.g., Kowalewski 1996).  
368 Further, such unstable substrates may have inhibited colonization of dead shells by encrusters, and their  
369 infestation by borers, thereby explaining the low degrees of taphonomic alteration for these processes

370 (Figure 14). However, possible causes for the observed differences in biogenic alteration (slightly higher  
371 in Waikaraka debris flow deposits for example) are multifarious, including high-sedimentation rates  
372 leading to rapid burial, unsuitable shell substrates and/or ecological conditions for infester communities  
373 (Caron et al. 2019).

374 Results of the taphonomic evaluation suggest that the alteration state of skeletal remains in the studied  
375 deposits not only reflects depositional and ecological conditions at the site of carbonate production on  
376 the shelf, but has also potential to record conditions of transport during their remobilization in gravity-  
377 driven systems. In particular, the taphonomy of skeletal contents, as a complementary tool to other  
378 analytical approaches, may provide clues on the laminar (*i.e.*, debris flows) versus turbulent (*i.e.*,  
379 turbidites) character of flow, and whether deposition by traction sedimentation prevailed over suspension  
380 sedimentation (Lowe 1982).

### 381 **6.3. Distinct substratum and continental connection**

#### 382 Lithoclast observations

383 At Homewood, the extra- and intraformational lithoclasts tend to oscillate between sub-angular to sub-  
384 rounded shapes and are in average of granule- to pebble-grades in the confined channel fills (**Fa2c**) and  
385 granule- to cobble-grades in the mass-wasting system (**Fa3p**). Conversely, at Waikaraka, sub-rounded  
386 clasts dominate and sub-angular shapes only preferentially occur in the early-stages confined channel  
387 fills (**Fa2c-e**). At Waikaraka, the clast grades also largely depend on the style of channel fills, and are  
388 preferentially of pebble- to outsized-grades in the early-stages fills (**Fa2c-e**) and of pebble-grades in the  
389 late-stage fills (**Fa2c-l**) (see Section 5).

390 Pre-subduction extraformational clasts largely dominate representing 95 to 99% of the overall lithoclast  
391 content at Homewood and 84% to 97% at Waikaraka (Figure 13). They mostly rework (>80%) Upper  
392 Cretaceous strata (*e.g.*, mainly siliceous and calcareous mudstones, rare lavas) and frequently include  
393 (~4 to 10%) Paleocene clasts (*e.g.*, limestones). Conversely, Lower Cretaceous clasts (*e.g.*, Torlesse  
394 greywackes) are only observed, sporadically (~2%), at Waikaraka, whereas Eocene clasts (*e.g.*, smectitic  
395 mudstones) are virtually absent and only rarely found (~1%) at Homewood (Figure 13).

396 At Waikaraka, the early-stage fills of the first channel complex (**Fa2c-e**) deviate from this model,  
397 remobilizing sub-angular to sub-rounded, pebbles to outsized clasts (deci- to decametric) from the entire  
398 Lower Cretaceous to Paleocene pre-subduction series at their base. Interestingly, sub-rounded to  
399 rounded, granule- to pebble-grade Lower Cretaceous clasts are also occasionally observed throughout

400 the confined slope channel system (**Fa2c**) (Figure 13). They are preferentially restricted to particular  
401 conglomerate intervals (**CF**, **gHDTc**) of the confined channel fills, generally located at the top of the early-  
402 stage fills (**Fa2c-e**).

403 The syn-subduction extraformational lithoclasts most commonly include pebble- to boulder-grade clasts  
404 of lithified shell beds. They are frequent in the confined slope channel system of Waikaraka. There, eight  
405 distinct varieties of shell bed clasts are observed (Appendix 5), which are often of cobble- to boulder-  
406 grades in the early-stage fills (**Fa2c-e**). Thick-shelled Molluscan remains are frequent in the shell bed  
407 clasts (Appendix 5, A-F, H), along with well-rounded lithic pebbles commonly bearing the trace fossil  
408 *Gastrochaenolites* (Appendix 5, B-D). The overall faunal assemblages, including *Struthiolaria* sp.,  
409 *Polinices* sp., Oysters, Turritellids, Serpulids and zooxanthellate corals (Appendix 5), and moderate  
410 degrees of pre-burial taphonomic alteration collectively point to rapid deposition in high-energy proximal  
411 inner-shelf to middle-shelf environments. Conversely, shell bed clasts are only punctually found (<0.5%)  
412 in the confined slope channel (**Fa2c**) and mass-wasting system (**Fa3p**) of Homewood. They however  
413 only comprise pebble- to cobble-grade skeletal and bioclastic fine-grained sandstones dominated by  
414 turritellid gastropods (cf. facies Fa6c from [Bailleul et al. \(2007\)](#)).

415 The remaining syn-subduction extraformational lithoclasts comprise pebble-grade wood fragments and  
416 disseminated plant debris. A few cobble- to boulder-grade clasts of fossil tubular concretions ([Malie et al.](#)  
417 [2017](#)) were also observed in the early-stage fills (**Fa2c-e**) at Waikaraka.

418 Finally, the syn-subduction intraformational lithoclasts are characterized by outsized, contorted  
419 sandstones and pebble- to boulder-grade mudstone clasts. The outsized, contorted sandstones are  
420 generally restricted to **DF**, particularly at Homewood; whereas the mudstone clasts, although present in  
421 some of the conglomerates, are dominantly found in the coarse-grained basal intervals of the sandstones  
422 at both Waikaraka and Homewood.

#### 423 Lithoclast interpretations

424 As previously observed elsewhere in the Coastal Ranges ([Neef 1992](#); [Bailleul et al. 2013](#); [Claussmann](#)  
425 [et al. 2021](#)), the nature of the extraformational lithoclasts informs that the mixed siliciclastic-carbonate  
426 shelfal domain sourcing the Waikaraka deposits developed above a pre- (e.g., Cretaceous to Paleogene  
427 series) and syn-subduction (e.g., Miocene sediments) substratum, whereas the paucity of  
428 extraformational syn-subduction clasts (e.g., a couple of shell bed clasts) at Homewood suggests a pre-  
429 subduction-dominated substratum.



430 At Waikaraka, the variety encountered in both the proportion and nature of pre-subduction material  
431 informs that the basal deposits of the confined slope channel system (**Fa2c**) remobilized the entire pre-  
432 subduction series down to the Lower Cretaceous, whereas the remainder mostly captured remobilized  
433 Upper Cretaceous strata (Figure 13). The absence of Lower Cretaceous greywackes and predominance  
434 of Upper Cretaceous strata at Homewood both suggest a slightly different substratum and or exhumed  
435 paleobathymetric high sourcing the deposits. The implications for both these localities are discussed in  
436 Section 7.

437 At Waikaraka, the syn-subduction material, mostly comprising shell bed clasts, likely results from  
438 destabilization of previously deposited and lithified strata (*i.e.*, substratum) rather than contemporaneous  
439 sediments. Their diversity captures different mixed siliciclastic-carbonate shelfal environments, which  
440 either (1) belonged to the same shelfal system, thus suggesting different phases of shelf development or  
441 local variations in depositional settings (Bailleul et al. 2015), and or (2) resulted from distinct shelfal  
442 systems that formed in the area since the onset of subduction (Bailleul et al. 2007; Bailleul et al. 2013;  
443 Caron et al. 2021). Conversely, at Homewood, the occasional sandy shell beds could be  
444 contemporaneous with the development of the shelfal environment. Early lithification of the deposits at  
445 or near the seafloor may have been triggered by increased saturation of seawaters with respect to  
446 CaCO<sub>3</sub>, following partial dissolution of aragonitic turritellid shells, though such process has been  
447 documented to predominantly result from burial diagenesis (*i.e.*, mesodiagenesis) (Nicolaidis 1995;  
448 Haywick 2004; Caron and Nelson 2009).

449 The abundant pebble-grade to outsized sub-angular clasts found in the confined slope channel (**Fa2c**)  
450 and mass-wasting system (**Fa3p**) of Homewood, and more sporadically in the early-stage confined  
451 channel fills (**Fa2c-e**) of Waikaraka, both indicate that despite clast interactions, not only the sediment  
452 sources were nearby but also that the material was “freshly” eroded to allow preservation of such features  
453 (Postma et al. 1988).

454 Notwithstanding partial remobilization of the shelf and slope substratum through submarine  
455 destabilization (*e.g.*, mass-wasting failure) and or erosional processes (*e.g.*, channel cutting through the  
456 slope substratum) to provide for most of the lithoclast content (*e.g.*, Sømme et al. 2009; Janocko et al.  
457 2013; Paumard et al. 2020), dismantlement and erosion of exhumed paleobathymetric highs probably  
458 contributed as well (*e.g.*, Ogata et al. 2019). In fact, the presence of wood fragments and plant debris in  
459 both settings informs that the source areas were connected to a terrestrial ecosystem (Kuenen 1964),  
460 which indicate the nearby presence and connection with exhumed land. Also, the occurrence of sub-  
461 rounded lithoclasts can suggest a fluvial and or shallow-marine origin with extraformational clasts that

462 resided for at least some time in a littoral zone, before being transferred to bathyal depths (Di Celma et  
463 al. 2010). Where grain collision is inferred as a dominant grain-support mechanism (e.g., in most of the  
464 **Fa2c** and **Fa3p** deposits), it can also be expected to have favored the quick abrasion and rounding of the  
465 harder, coarse-grained clasts.

466 The large volume of wood fragments and plant debris recorded throughout the confined slope channel  
467 deposits of Waikaraka (**Fa2c**) and the well-developed turbidite system of Homewood (**Fa1g**) both support  
468 source connections to mature hinterlands (e.g., Stanley 1986; McArthur et al. 2016b). The discrete  
469 occurrences observed in the mass-wasting system of Homewood (**Fa3p**) rather points toward a small  
470 and or isolated source system, such as an island developed above tectonically-controlled topography  
471 (McArthur et al. 2016a), hence reinforcing the hypothesis of a source system detached from the main  
472 shelf domain.

473 Finally, the nature of intraformational clasts, such as mud clasts, informs that these gravity flows were  
474 erosive and ploughed through mud-prone substrates as they moved downslope or at hydraulic jumps  
475 (Posamentier and Martinsen 2011). At Homewood, the contorted sandstones also indicate of the varying  
476 nature of the ploughed substrate, locally supported by the underlying sandier deposits (**Fa2c**).

## 477 7. DISCUSSION

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### 478 7.1. Gravity-driven systems from forelimb, continent-attached shelfal 479 source

480 The depositional system proposed for Waikaraka is one of a confined slope channel system (CSCS)  
481 (*sensu* Campion et al. 2000) distributing land-derived, shallow- and deep-marine bedrock-derived  
482 sediments to the trench-slope basin.

483 As interpreted (see Sections 5 and 6), this system was likely sourced from the regional, narrow and  
484 continent-attached mixed siliciclastic-carbonate shelfal domain, which was contemporaneously  
485 developing to the west during the Middle Miocene, atop seaward-verging thrust-cored anticlines (Figure  
486 15A; Figure 16) (Crundwell 1987; Chanier 1991; Bailleul et al. 2007; Bailleul et al. 2013; Bailleul et al.  
487 2015; Clausmann et al. 2021).

488 Precise geometry of the slope upon which the CSCS developed is here a matter of speculation; yet, the  
489 forelimb settings rather suggest a steep, probably irregular, tectonically-controlled gradient close to the

490 shelf edge and across the slope (Figure 15A; Figure 16) (e.g., Naranjo-Vesga et al. 2020). The large  
491 volume of terrigenous material as well as the nature of the reworked clasts (e.g., well-preserved bioclasts,  
492 shell fragments, pre-subduction strata) recorded throughout the CSCS deposits (**Fa2c**) both indicate  
493 persistent connection with a shelfal domain, recurrently supplied by the mature, vegetated hinterland to  
494 the west, exhumed since the onset of the Hikurangi subduction and mostly exposing Cretaceous strata  
495 (Figure 16) (Chanier and Ferrière 1991).

496 Along active margins, CSCS commonly represents long-lived transport routes for sediments across the  
497 shelf and slope that are typically insusceptible to sea-level changes and directly connect the hinterland  
498 (Type I, shelf-incising and river-associated system of Harris and Whiteway (2011)) and or shoreface  
499 depositional systems (Type II, shelf-incising system of Harris and Whiteway (2011)) to the deep sea  
500 (Sømme et al. 2009). Here however, although regional tectonics (i.e., margin uplift) largely governed the  
501 initial emplacement of this mixed siliciclastic-carbonate shelfal domain installed atop thrust-cored  
502 anticlines (Bailleul et al. 2013), eustatic processes appear to then chiefly control their development and  
503 stratigraphic architecture (Bailleul et al. 2015). Notwithstanding the reworking of Upper Cretaceous  
504 conglomerates characteristically rich in already-reworked Lower Cretaceous Torlesse greywackes  
505 (Chanier and Ferrière 1991) as a potential source for providing gravels readily available to be transferred  
506 across the narrow shelf, the punctual input of well-rounded pebbles of Torlesse greywackes observed at  
507 the top of the early-stages fills of the CSCS (see Section 6) could also be attributed to an intermittent  
508 connection of the shelf with river(s). The successive T-R sequences that characterize these shelfal  
509 environments (Bailleul et al. 2015) could explain such periodic communication with the exhumed  
510 Cretaceous hinterland and would in turn suggest some eustatic control here, eventually captured in the  
511 CSCS infill.

512 The broad fining- and thinning-upwards trends observed in the CSCS (**Fa2c**) are common in many other  
513 slope channel systems, indicating channel bypass and backfilling processes (e.g., Mutti 1985; Deptuck  
514 et al. 2003; Beaubouef 2004; Sylvester et al. 2011; Kneller et al. 2020). The repeating series of  
515 architecture and facies associations inform that this CSCS system likely behaved in a cyclic manner  
516 through time that may well have been in phase with the sea-level changes, yet likely also resulted from  
517 external forcing factors, such as pulses of thrust fault activity (e.g., uplift and or related seismicity) at the  
518 edge of the shelf, known to periodically renew erosion and active sediment transport into CSCS along  
519 active margins (e.g., Mountjoy et al. 2009; Mountjoy et al. 2018; Watson et al. 2020).

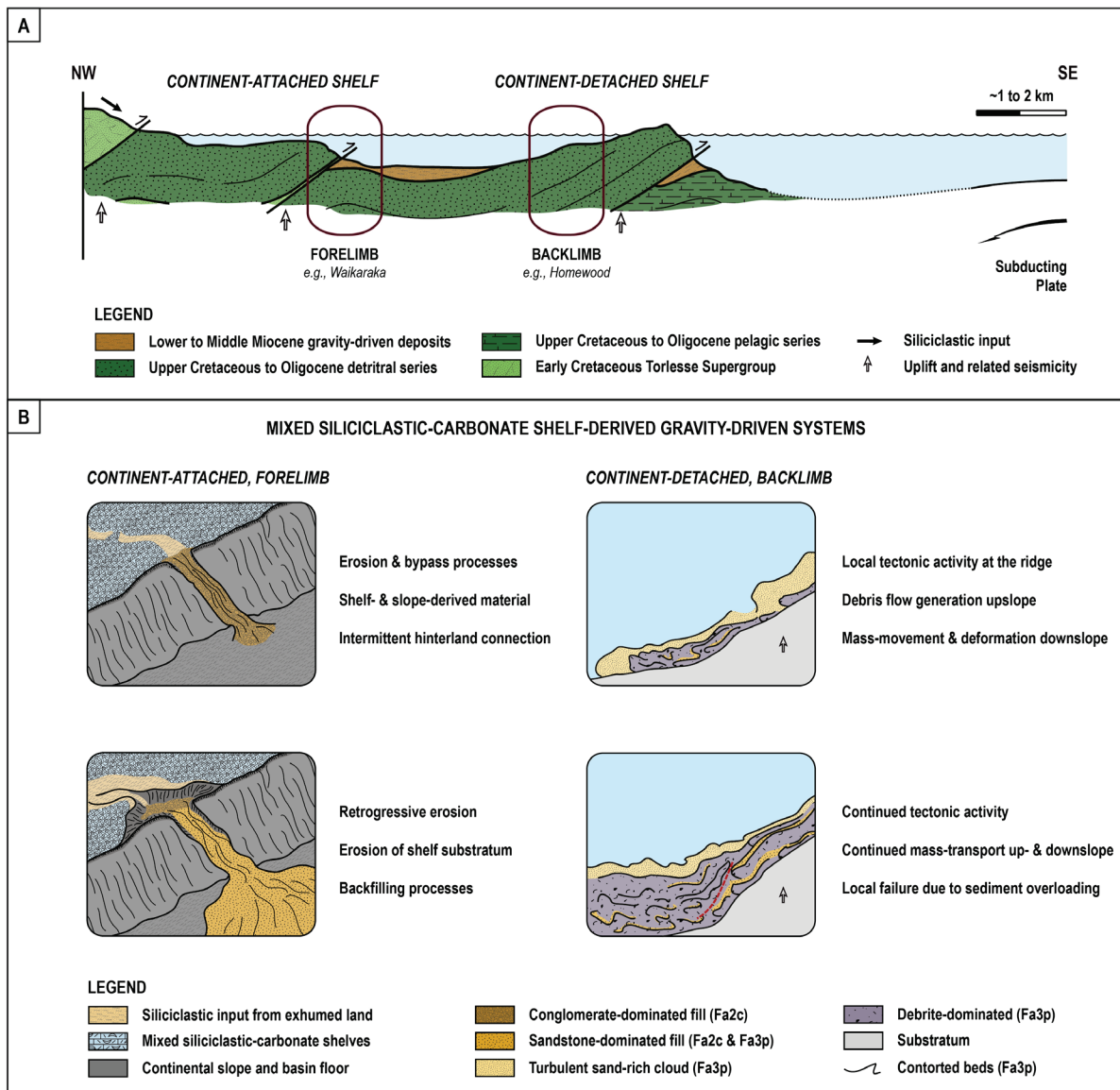
520 Interestingly, the reworked material contradicts the traditional unroofing model, incorporating clasts from  
521 the entire Lower Cretaceous to Paleogene pre-subduction strata only within its lowermost record and



522 being otherwise dominated by Upper Cretaceous strata (see Section 6.3). Although direct erosion of the  
523 exhumed Cretaceous basement likely acted as an additional source, we attribute the former clast content  
524 to chiefly result from CSCS forming processes, such as initial slope failure (e.g., Janocko et al. 2013) and  
525 or slope erosion (e.g., Sømme et al. 2009). These processes exposed the underlying submarine bedrock  
526 of the thrust-cored anticline (i.e., forelimb setting) upon which the shelfal domain developed, making it  
527 ready to be included into the subsequent gravity flows (Figure 15A; Figure 16) (e.g., Ogata et al. 2019).  
528 For the latter contrasting clast content, we here infer the autogenic evolution of this shelf-incising CSCS  
529 (Type I, shelf-incising and river-associated system of Harris and Whiteway (2011)) during late low-stand  
530 conditions, whereby retrogressive erosion at the head of the system (*sensu* Daly 1936) would have  
531 favored the failure, storage and reworking of the marginal siliciclastic-carbonate sediments,  
532 contemporaneously (e.g., living or freshly deceased macrofauna, terrigenous debris) and previously  
533 deposited (e.g., shell bed lithoclasts), along with their underlying substratum, here dominated by Upper  
534 Cretaceous strata (Figure 15B; Figure 16). The upward increase of bioclastic material recorded in the  
535 channel fill may also partially illustrate such progressive upslope variation in source region, whereas the  
536 dominance of skeletal grains (over non-skeletal, e.g., ooids, pellets) typically occur during such sea-level  
537 low-stands (Reijmer et al. 2015).

538 Here, regional tectonics appear to have chiefly governed both the birth (emplacement) and death  
539 (drowning) of (1) the continent-attached mixed siliciclastic-carbonate shelfal domain installed atop thrust-  
540 cored anticlines (Bailleul et al. 2013; Bailleul et al. 2015) and of (2) its related shelf-derived, deep-marine  
541 systems. Whilst the shelfal domain development then became largely eustatically-controlled (Bailleul et  
542 al. 2015), the combination of steep slope gradient, high sediment supply and narrow shelf width resulted  
543 in the development of a CSCS (**Fa2c**) that efficiently transferred the shelf-derived sediments to the deep-  
544 marine environments (e.g., Naranjo-Vesga et al. 2020). Although punctually capturing the sea-level  
545 changes, the overall evolution and geometry infill of the CSCS seemed essentially controlled by local  
546 tectonic activity at the mixed siliciclastic-carbonate shelf edge, which, in turn, determined its stratigraphic  
547 infill (sediment distribution and architecture) (Figure 15; Figure 16).

548 The abrupt changes in sedimentation styles (i.e., sand-dominated) and paleodepth of deposition (i.e.,  
549 lower bathyal) recorded in the uppermost CSCS deposits (see Section 5; Appendix 2) could be attributed  
550 to the major change in tectonic regime (i.e., generalized subsidence) that was recorded along the  
551 Hikurangi margin at the time, which in turn resulted in rapid bathymetric deepening of the area (i.e., high-  
552 stand conditions) (Chanier et al. 1999; D5 discontinuity and U5 unit of Bailleul et al. 2013) (Figure 10a,  
553 e). Direct coarse-grained sediment supply stopped; yet, some high-stand activity was maintained along  
554 the CSCS (e.g., Mountjoy et al. 2009).



555

556 Figure 15: (A): Schematic NW-SE cross-section through the subduction-related, compressive Whareama intra-slope basin

557 (*i.e.*, trench-slope basin) during the Middle Miocene. The depositional settings for the forelimb and backlimb settings of actively

558 growing thrust-cored structures are highlighted. (B): Schematic representation of the inferred shelf-derived gravity-driven

559 systems and processes related to the forelimb and backlimb settings.

560

## 561 7.2. Gravity-driven systems from backlimb, continent-detached shelfal 562 source

563 The source and depositional system envisaged at Homewood contrast with the coeval continent-attached

564 shelfal environment and slope channel delivery system of Waikaraka. As previously described (see

565 Sections 5 and 6), the nature of the deposits (**Fa2c** and **Fa3p**) indicates a shelfal environment that

566 developed upon and around a structurally-controlled ridge located to the east of Homewood, which was  
567 detached from the regional shelf and hinterland (Figure 15A).

568 Isolated platforms may develop above fault-growing folds and favor biogenic carbonate production as  
569 they lack continent-derived siliciclastic inputs (Caron et al. 2004; Ćosović et al. 2018; Caron et al. 2021).  
570 At Homewood however, analyses of the reworked material (see Section 6) inform that mixed siliciclastic-  
571 carbonate sediments dominated, thereby suggesting both emergence and subsequent erosion of the  
572 uplifted ridge (Figure 15A; Figure 16). Previous work highlighted that islands developed above  
573 tectonically-controlled topography can supply terrigenous material (e.g., McArthur et al. 2016a) and result  
574 in mixed siliciclastic-carbonate sedimentation (e.g., Chiarella et al. 2012; Chiarella and Longhitano 2012;  
575 Ćosović et al. 2018).

576 More particularly, the nature of the deposits (i.e., **Fa2c** and **Fa3p** slope deposits) and their clasts (see  
577 Part 5) suggests that the ridge was nearby and exhumed for some time to allow (1) the dismantlement of  
578 the pre-subduction series down to the Upper Cretaceous strata, (2) their subsequent reworking in the  
579 littoral domain and (3) the formation of sandy shell beds. Notwithstanding submarine erosion, the  
580 abundance of sub-rounded clasts of Upper Cretaceous strata implies that these strata likely outcropped  
581 at the time (e.g., Di Celma et al. 2010), thereby inferring initial rise and emergence of the ridge prior to  
582 the Middle Miocene, Upper Clifdenian to Lower Lillburnian (Langhian, 15.9 – 14 Ma, Appendix 2).

583 When combined, the pre-subduction Paleogene cover, about 600 meters thick in the study area (Van den  
584 Heuvel 1960; Chanier 1991; Chanier and Ferrière 1991), and the general exhumation rate of up to 0.3  
585 kilometer/million year calculated by Jiao et al. (2014) along the eastern North Island of New Zealand  
586 indicate that the ridge emergence (i.e., island setting) and subsequent erosion started at least ca. 2 Ma  
587 prior to that time. This ridge likely corresponded to the seaward margin of the Whareama Basin,  
588 structurally controlled by the underlying seaward-verging Flat Point-Whakataki Fault complex and located  
589 to the east of Homewood (i.e., Glenburn Nappe) (e.g., Chanier 1991) (Figure 3; Figure 15A). The set of  
590 morphometric relationship equations calculated by Moscardelli and Wood (2015) predicts a length of  
591 about two kilometers for the MTDs (Appendix 6), which may or may not include the evacuation area  
592 (*sensu* Moscardelli and Wood 2015) yet provides a broad insight as to the potential distance to the ridge  
593 at the time (less than two kilometers).

594 The envisaged shelfal system was narrow and located on the western side of the island, upon the  
595 backlimb of the underlying asymmetrical thrust (Figure 15A; Figure 16). Similarly, the mixed siliciclastic-  
596 carbonate Fingerpost shelf developed further north during the Early to Middle Miocene; it was also  
597 positioned to the west of an emergent landmass (i.e., the Cape Turnagain Structural High), itself



598 controlled by an underlying seaward-verging thrust (*i.e.*, backlimb setting) (Bailleul et al. 2013, their figure  
599 12).

600 At Homewood, the shelf was likely cut by a network of small slope channels connecting the island through  
601 the shelf to the Whareama Basin (**Fa2c**) (Figure 16). The transition from slope channel (**Fa2c**) to shelf-  
602 derived mass-wasting (**Fa3p**) systems is attributed to a period of tectonic activity (*e.g.*, uplift) at the basin-  
603 bounding structure, resulting in growth strata deposition on the eastward-margin of the basin (see Section  
604 5.3) (*e.g.*, Moore and Karig 1976; Stevens and Moore 1985) and typically lasting 1 to 2 Ma in these  
605 settings (Nicol et al. 2002; Bailleul et al. 2013).

606 Structure growth both (1) promoted the repeated destabilization of the isolated shelf and generation of  
607 the **Fa3p** mass-wasting flows as well as (2) controlled the axial routing and syn-kinematic development  
608 of the overlying **Fa1g** turbidite lobe system sourced from the mature hinterland to the west (Figure 16).  
609 The associated MTDs (**Fa3p**), only ~15 meters thick, recorded a difference of ~40° in bed dips (*i.e.*,  
610 growth strata), thereby indicating complex interplay between the growing structure and sediment  
611 deposition. The slope gradient acted as a dynamic control triggering (1) mass-wasting processes  
612 upslope, as well as, down the slope, (2a) rapid flow transformation and efficient dilution (*e.g.*, Fisher 1983;  
613 Strachan 2008) resulting into fast-moving multiphase debris flows (*e.g.*, Postma 1986; Mohrig et al. 1998;  
614 Sohn et al. 2002) and (2b) coeval remobilization (soft-sediment translation and deformation) of the  
615 previously deposited, unlithified mass-flow deposits (*e.g.*, Maltman 1994) (Figure 15B; Figure 16).

616 The MTDs' succession and the variations in slope gradient recorded throughout can therefore be used  
617 as a proxy for the tectonic activity of an individual structure. Here, they inform that at least six main  
618 tectonic pulses with uplift affected the structurally-controlled shelf-margin(s), each respectively  
619 generating one to several coalescing debris flows. Towards the top of the MTDs' succession, the high  
620 amount of sheared debris flow deposits, contorted strata, and the increasing bed dip variations (Figure  
621 9) suggest both steep slope gradients (*e.g.*, Postma 1986) and an acceleration of the ridge activity. Steep,  
622 rising slopes, recurrently loaded with sediments, are unstable (Hein and Gorsline 1981) and thus likely  
623 provided the right settings for locally triggering the failure of the entire MTDs' series as previously  
624 observed at Homewood (Figure 9; Figure 15).

625

## 7.3. Generic implications

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### 7.3.1. For shelf-derived gravity-driven systems sourced from thrust forelimb

627

#### and backlimb

628

In this study, the high variability encountered within one intra-slope basin in terms of sediment delivery systems as well as transport and depositional processes highlights the role and importance of varying structural settings of the sediment source for controlling the subsequent deep-marine depositional settings. Despite similar source environments (*i.e.*, mixed siliciclastic-carbonate shelves) developed atop the basin-bounding structures (*i.e.*, thrust) along with comparable sediment delivery systems (*i.e.*, slope channel and remobilization) and nature of flow processes (*i.e.*, gravity-driven), the resulting shelf-derived deposits downslope are characterized by contrasting lithofacies assemblages, sizes and geometries.

635

Notwithstanding the size and extent of the catchment area or continental connection (attached or detached) as critical controls (Sømme et al. 2009), we here suggest that along active margins, the geometry and activity of the underlying structures (thrust forelimb and backlimb settings) at the shelf edges also greatly influence the resulting sediment distribution and architecture.

639

The shelf-edges on the forelimbs of underlying thrusts are abrupt, generally associated with steep, irregular and tectonically-controlled slope profiles (Naranjo-Vesga et al. 2020). Erosion and transport capacities proportionally increase with slope gradient and thus favor significant sediment bypass to the deeper settings (Sømme et al. 2009; Crisóstomo-Figueroa et al. 2020). Here, the tectonic activity at the shelf edges generally favored the development of systems that dynamically interact and rejuvenate the slopes (*e.g.*, Waikaraka retrogressive system or Sefton Hills repeated collapse events (Clausmann et al. 2021)) and the resulting gravity-driven systems are generally hundreds of meters thick and several kilometers wide, preferentially transporting sediments through sediment gravity flows.

647

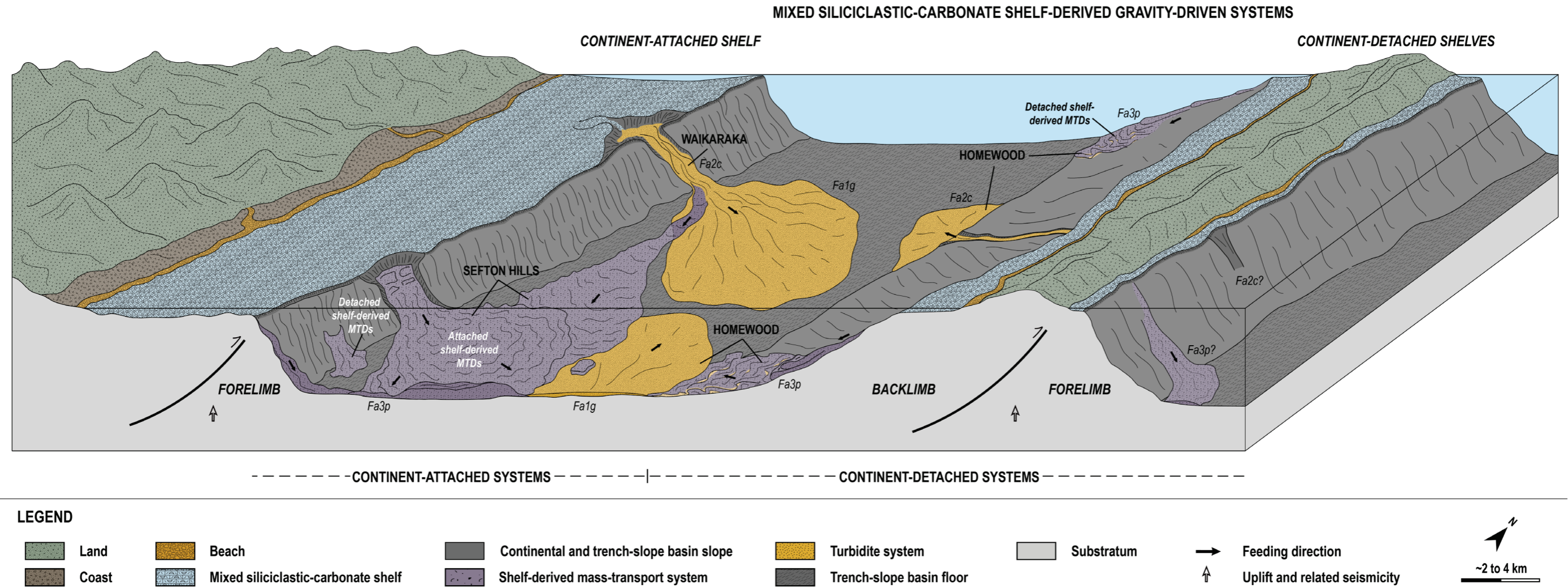
Conversely, thrust backlimb commonly provide longer and gentler slope gradients upon which a more gradual shelf edge transition may be expected and may facilitate the development of less efficient depositional systems (Sømme et al. 2009). In this study however, these settings were connected to limited catchment areas (*e.g.*, isolated island), thereby preventing the development of particularly large systems that typically occur on gentler slopes. Therefore, despite similar delivery systems to those associated to forelimb-related slopes (*e.g.*, slope channels and destabilization), their sizes were largely smaller (tens of meters thick).



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Figure 16: Contrasting depositional systems for mixed siliciclastic-carbonate shelf-derived gravity-driven flows along tectonically active margins, using the southern portion of the Hikurangi subduction margin as an example (Whareama trench-slope basin). Fa1g – refers to depositional lobe system; Fa2c – to confined slope channel system; Fa3p – to shelf-derived mass-wasting systems. Depositional settings from the Sefton Hills localities were adapted from Clausmann et al. (2021).



660 Destabilization and collapses of the mixed siliciclastic-carbonate shelfal sources occurred on both sides  
661 of the basin-bounding structures and thus indicate that any of these shelves, either attached to the main  
662 continental domain or detached from it (e.g., isolated island) can be subjected to shelf-derived sediment  
663 mass movement and mobilization (*sensu Iverson 1997*).

664 When sourced from the backlimb of an actively growing structure (i.e., thrust) however, the mass-wasting  
665 deposits exhibit more complex internal architectures, clearly recording the slope gradient variations and  
666 resulting in intricate interactions between mass movements and mass flows, such as the ones observed  
667 at Homewood. The changes in slope gradient will dynamically control both (1) the generation of mass  
668 movements and flows upslope, and (2) the coeval remobilization (e.g., sliding) and soft-deformation (e.g.,  
669 slumping) of the previously deposited, poorly-lithified sediments (e.g., debrites) downslope. Conversely,  
670 although mass movement (e.g., slide, slump) can undoubtedly occur as well, the MTDs sourced from the  
671 forelimb of an actively growing structure (i.e., thrust) seem primarily characterized by a series of  
672 coalescing mass-flow deposits (e.g., debris flows) (e.g., Festa et al. 2010 and references therein). Mass  
673 movements either seem preferentially recorded close to the source region or at the base of the MTDs'  
674 sequence, where they often result from soft-sediment deformations (e.g., slide, slump) of the unlithified  
675 sediments that were previously deposited at the front of the rising structure (e.g., Festa et al. 2010; Lackey  
676 et al. 2020).

677 Overall, the shelf morphology (i.e., narrow), extent (e.g., regional or isolated), continental connection  
678 (e.g., attached or detached), sediment type (i.e., mixed siliciclastic-carbonate sedimentation) as well as  
679 the respective geotectonic settings (e.g., upon and around the forelimb or backlimb of the underlying  
680 thrusts and fault-growing folds) all chiefly controlled the architecture and subsequent shelf-derived  
681 deposits. However, along tectonically-active margins, when sourced from mixed siliciclastic-carbonate  
682 systems, the deep-marine, shelf-derived deposits seem characterized by tens to a few hundred of meters  
683 of thickness and several kilometers of lateral extent. The durations of the associated depositional systems  
684 are from 1 to 2 Ma, being controlled by the structural styles and local tectonic activity at the source rather  
685 than regional-scale variations.

### 686 **7.3.2. For mass-transport deposit nomenclature**

687 Moscardelli and Wood (2015) have proposed differentiating mass-wasting products into *attached* and  
688 *detached* systems, respectively sitting in a proximal and a distal position from any shelf break, and either  
689 sourced from this regional area (e.g., shelf) or local structures. *Attached* MTDs commonly represent the  
690 largest occurrences, being regionally extensive, and are essentially triggered by sea-level variations and

691 or high sedimentation rates (Moscardelli and Wood 2008; Posamentier and Martinsen 2011; Moscardelli  
692 and Wood 2015). As discussed here however, along active margins, shelfal environments not only  
693 develop in the submerged prolongation of the main hinterland, constituting continent-attached shelves,  
694 which can source large-scale shelf-derived mass-wasting products (e.g., Claussmann et al. 2021). Shelfal  
695 environments can also form episodically upon and around structurally controlled and active ridges,  
696 resulting in shelves that are detached from the main shelf domain (e.g., Ćosović et al. 2018). These  
697 isolated settings, with a virtually absent to limited catchment area (whether underwater or emerged),  
698 feature reduced shelfal source systems, which will, in turn, produce significantly smaller shelf-derived  
699 mass-wasting products such as the ones described at Homewood. The size of the resulting deposits best  
700 compares to those commonly found in *detached* MTDs (*sensu* Moscardelli and Wood 2015) (Appendix  
701 6) whereas their internal characteristics inform that tectonic acted as the main triggering mechanism and  
702 that the source was nearby.

703 Notwithstanding sea-level changes and high sedimentation rates as causal mechanisms for shelf-derived  
704 mass-wasting events (Posamentier and Walker 2006; Moscardelli and Wood 2008; Moscardelli and  
705 Wood 2015; and references within), Bailleul et al. (2013) and Claussmann et al. (2021) previously  
706 demonstrated that along active margins, periods of tectonic activity (ca. 1 to 2 Ma) constitute another  
707 mechanism to consider for both the development and subsequent shelf failures. The MTDs described in  
708 this study not only provide another example of such shelf-derived mass-wasting products, whereby  
709 tectonic activity controlled the repeated destabilization and collapses of the shelfal source (see also  
710 Romero-Otero et al. 2010; Ortiz-Karpf et al. 2018), but also indicate that shelf-derived mass-wasting  
711 events do not always result in large-scale systems (Appendix 6) (see also Claussmann et al. 2021).  
712 Similarly, Naranjo-Vesga et al. (2020) observed that owing to the complex relationships that exist  
713 between the shelf width, slope profile and sediment supply in an analogous convergent setting (*i.e.*,  
714 southern Caribbean fold-and-thrust belt), both *attached* and *detached* MTDs could be sourced from the  
715 regional shelf and slope regions.

716 Therefore, we here propose to slightly amend the MTD nomenclature from Moscardelli and Wood (2008)  
717 and Moscardelli and Wood (2015) by using the related '*attached*' and '*detached*' morphometry  
718 characteristics independently from the type (e.g., shelf, slope, local bathymetric highs) and distance from  
719 source areas to better account for the variety of geotectonic settings (e.g., convergent).

720 As a result, we introduce the term '*shelf-derived* systems' which refers to all the MTDs that result from  
721 the destabilization and collapses of shelfal environments, whether induced by regional and or local causal  
722 mechanisms (sea-level changes, high sedimentation rates, periods of repeated tectonic activity, etc.),

723 and which can either source large-scale '*attached shelf-derived systems*' or smaller '*detached shelf-*  
724 *derived systems*' (*sensu* Moscardelli and Wood 2008; Moscardelli and Wood 2015) (Figure 16). Similar  
725 reasoning can be applied to the MTDs that result from failures of the regional slope (*i.e.*, '*slope-derived*  
726 *systems*').

727 Notwithstanding the slope profile, here inherently linked to the geometry of the underlying structural styles  
728 at the source (*e.g.*, forelimb or backlimb) and the sediment supply as other critical parameters (*e.g.*,  
729 Naranjo-Vesga et al. 2020), the present study demonstrates that the geomorphology of source area, and  
730 more particularly its connection to the continent (*i.e.*, catchment area) is also critical. Here, the narrow  
731 shelves appear to source both '*attached and detached shelf-derived systems*' when connected to the  
732 main hinterland (narrow continent-attached shelf) and '*detached shelf-derived systems*' when isolated  
733 and disconnected from it (narrow continent-detached shelf) (Figure 16). Further work is however required  
734 to generalize such an observation; and more particularly, the impact of the forelimb settings on the  
735 resulting *shelf-* and *slope-derived* mass-wasting systems (*e.g.*, size, internal characteristics, causal  
736 mechanisms) sourced from continent-detached mixed siliciclastic-carbonate shelfal domains remains to  
737 be investigated.



## 738 8. CONCLUSIONS

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739 In this study, we used high-resolution outcrop data to investigate the implications of mixed siliciclastic-  
740 carbonate source systems developed upon and around the forelimbs and backlimbs of actively growing  
741 asymmetrical structures for the development of the subsequent gravity-driven systems beyond the shelf  
742 edges. Particularly, this study highlights the role and importance of varying structural setting of the  
743 sediment source in controlling the morphologies of the delivery systems, the patterns of sediment  
744 dispersal, and the related sedimentary facies along active margins.

745 First, we demonstrated that during the Middle Miocene, the paleogeography of the southern, inner portion  
746 of the Hikurangi Margin was diversified, comprising a mature hinterland to the west as well as isolated  
747 islands to the east, developing atop the actively growing seaward-verging thrusts of the subduction  
748 wedge. Mixed compositional siliciclastic-carbonate sedimentation dominated at shallow depths and the  
749 different morphologies of the source regions (continent-attached, forelimb and continent-detached,  
750 backlimb) resulted in the development of highly varied shelf-derived gravity-driven systems downslope  
751 within very short distances (less than five kilometers) that interacted with the structures and were  
752 eventually captured both on their margins and within the trench-slope basins.

753 Then, we showed that mixed siliciclastic-carbonate gravity-driven systems in active settings appear  
754 preferentially characterized by deposition of tens to a few hundred of meters of thick units, displaying  
755 several kilometers of lateral extent, for duration of ca. 1 to 2 Ma. Such systems are primarily controlled  
756 by the geometries and tectonic motion of the underlying structures at the shelf edges. Different  
757 morphologies of the source regions (*i.e.*, the aforementioned continent-attached forelimb and continent-  
758 detached backlimb of thrust settings) result in the development of distinct paleoenvironmental conditions  
759 at the shelf and contrasting shelf-derived gravity-driven systems along and downslope. The shelf-edges  
760 on the forelimbs of actively growing thrusts are abrupt and commonly associated with irregular steep  
761 slopes. Here, they appear to have favored the development of systems that dynamically rejuvenate the  
762 slopes (*e.g.*, Waikaraka retrogressive systems). On the backlimbs however, the shelf-edges are generally  
763 associated to gentler slopes; yet, pulses in tectonic activity will increase slope gradient variations, thus  
764 promoting the development of sedimentary systems that directly interact with the slope (*e.g.*, Homewood  
765 mass-wasting system).

766 Finally, this study highlights that, whether they are caused by regional and or local triggering mechanisms,  
767 the destabilization and collapses of shelfal environments source '*shelf-derived* mass-wasting systems',  
768 characterized by two types: (1) large-scale, regional '*attached shelf-derived systems*' and (2) small-scale,

769 localized '*detached shelf-derived systems*'. Along active margins, and particularly the Hikurangi Margin,  
770 the '*shelf-derived mass-wasting systems*' are sourced from shelves that are either attached to the main  
771 continental domain or detached from it (e.g., isolated islands). The associated *shelf-derived* mass-  
772 wasting deposits can be found on both sides of actively growing asymmetrical thrusts. When sourced  
773 from the backlimbs however, the subsequent shelf-derived MTDs exhibit more complex internal  
774 architectures, ultimately recording the dynamic changes in slope gradient, which can both trigger the  
775 generation of mass failures and mass flows upslope as well as the coeval remobilization (translation and  
776 deformation) of the previously deposited MTDs downslope. As a result, the MTDs from thrust backlimbs  
777 can be used as proxies for unraveling the tectonic activity of an individual structure.

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- Abreu V, Sullivan M, Pirmez C, Mohrig D. 2003. Lateral accretion packages (LAPs): an important reservoir element in deep water sinuous channels. *Marine and Petroleum Geology*. 20(6):631–648.  
<https://doi.org/10.1016/j.marpetgeo.2003.08.003>
- Alsop GI, Weinberger R. 2020. Are slump folds reliable indicators of downslope flow in recent mass transport deposits? *Journal of Structural Geology*. 135:104037.  
<https://doi.org/10.1016/j.jsg.2020.104037>
- Alsop GI, Weinberger R, Marco S, Levi T. 2019. Identifying soft-sediment deformation in rocks. *Journal of Structural Geology*. 125:248–255.  
<https://doi.org/10.1016/j.jsg.2017.09.001>
- Armitage DA, Romans BW, Covault JA, Graham SA. 2009. The Influence of Mass-Transport-Deposit Surface Topography on the Evolution of Turbidite Architecture: The Sierra Contreras, Tres Pasos Formation (Cretaceous), Southern Chile. *Journal of Sedimentary Research*. 79(5):287–301.  
<https://doi.org/10.2110/jsr.2009.035>
- Bailleul J, Caron V, Chanier F, Mahieux G, Malié P, Gagnaison C, Potel S, Ferrière J. 2015. Modalités de la sédimentation de plate-forme sur la pente d'un prisme d'accrétion: l'exemple du Miocène Moyen du prisme Hikurangi (Nouvelle-Zélande). In: *Proceedings of the Congrès Français de Sédimentologie 15*. Chambéry, France; p. 47.
- Bailleul J, Chanier F, Ferrière J, Robin C, Nicol A, Mahieux G, Gorini C, Caron V. 2013. Neogene evolution of lower trench-slope basins and wedge development in the central Hikurangi subduction margin, New Zealand. *Tectonophysics*. 591:152–174.  
<https://doi.org/10.1016/j.tecto.2013.01.003>
- Bailleul J, Robin C, Chanier F, Guillocheau F, Field B, Ferrière J. 2007. Turbidite Systems in the Inner Forearc Domain of the Hikurangi Convergent Margin (New Zealand): New Constraints on the Development of Trench-Slope Basins. *Journal of Sedimentary Research*. 77(4):263–283.  
<https://doi.org/10.2110/jsr.2007.028>
- Ballance PF. 1976. Evolution of the Upper Cenozoic Magmatic Arc and plate boundary in northern New Zealand. *Earth and Planetary Science Letters*. 28(3):356–370.  
[https://doi.org/10.1016/0012-821X\(76\)90197-7](https://doi.org/10.1016/0012-821X(76)90197-7)
- Beaubouef RT. 2004. Deep-water leveed-channel complexes of the Cerro Toro Formation, Upper Cretaceous, southern Chile. *AAPG Bulletin*. 88(11):1471–1500.  
<https://doi.org/10.1306/06210403130>
- Beavan J, Tregoning P, Bevis M, Kato T, Meertens C. 2002. Motion and rigidity of the Pacific Plate and implications for plate boundary deformation. *Journal of Geophysical Research: Solid Earth*. 107(B10):ETG 19-1-ETG 19-15.  
<https://doi.org/10.1029/2001JB000282>
- Best MMR, Kidwell SM. 2000. Bivalve taphonomy in tropical mixed siliciclastic-carbonate settings. I. Environmental variation in shell condition. *pbio*. 26(1):80–102.  
[https://doi.org/10.1666/0094-8373\(2000\)026<0080:BTITMS>2.0.CO;2](https://doi.org/10.1666/0094-8373(2000)026<0080:BTITMS>2.0.CO;2)
- Best MMR, Ku TCW, Kidwell SM, Walter LM. 2007. Carbonate Preservation in Shallow Marine Environments: Unexpected Role of Tropical Siliciclastics. *The Journal of Geology*. 115(4):437–456. <https://doi.org/10.1086/518051>
- Bland KJ, Uruski CI, Isaac MJ. 2015. Pegasus Basin, eastern New Zealand: A stratigraphic record of subsidence and subduction, ancient and modern. *New Zealand Journal of Geology and Geophysics*. 58(4):319–343.  
<https://doi.org/10.1080/00288306.2015.1076862>



- Bouma AH. 1962. Sedimentology of some Flysch deposits: a graphic approach to facies interpretation [PhD Thesis]. Amsterdam, Netherlands: Elsevier.
- Bradshaw JD. 1989. Cretaceous geotectonic patterns in the New Zealand Region. *Tectonics*. 8(4):803–820.  
<https://doi.org/10.1029/TC008i004p00803>
- Browne GH, Bull S, Annot M, Boyes AF, King PR, Helle K. 2020. The role of mass transport deposits contributing to fluid escape: Neogene outcrop and seismic examples from north Taranaki, New Zealand. *Geo-Marine Letters*. 40:789–807. <https://doi.org/10.1007/s00367-020-00641-z>
- Brunt RL, Hodgson DM, Flint SS, Pringle JK, Di Celma C, Pr lat A, Grecula M. 2013. Confined to unconfined: Anatomy of a base of slope succession, Karoo Basin, South Africa. *Marine and Petroleum Geology*. 41:206–221.  
<https://doi.org/10.1016/j.marpetgeo.2012.02.007>
- Bull S, Cartwright J, Huuse M. 2009. A review of kinematic indicators from mass-transport complexes using 3D seismic data. *Marine and Petroleum Geology*. 26:1132–1151.
- Burgreen B, Graham S. 2014. Evolution of a deep-water lobe system in the Neogene trench-slope setting of the East Coast Basin, New Zealand: Lobe stratigraphy and architecture in a weakly confined basin configuration. *Marine and Petroleum Geology*. 54:1–22.  
<https://doi.org/10.1016/j.marpetgeo.2014.02.011>
- Calhoun NC, Clague JJ. 2018. Distinguishing between debris flows and hyperconcentrated flows: an example from the eastern Swiss Alps. *Earth Surface Processes and Landforms*. 43(6):1280–1294.  
<https://doi.org/10.1002/esp.4313>
- Campion KM, Sprague AR, Mohrig D, Lovell RW, Drzewiecki PA, Sullivan MD, Ardill JA, Jensen GN, Sickafoose DK. 2000. Outcrop Expression of Confined Channel Complexes. In: *Deep-Water Reservoirs of the World: 20th Annual*. Vol. 44. Oklahoma, America: SEPM (Society for Sedimentary Geology); p. 127–150.
- Caron V. 2011. Contrasted textural and taphonomic properties of high-energy wave deposits cemented in beachrocks (St. Bartholomew Island, French West Indies). *Sedimentary Geology*. 237(3):189–208.  
<https://doi.org/10.1016/j.sedgeo.2011.03.002>
- Caron V, Bailleul J, Chanier F, Mahieux G. 2021. Episodes of seabed rise and rapid drowning controlling the development of regressive and transgressive rhodolitic limestones in a tectonically-active subduction setting (Early Miocene, Wairarapa region, New Zealand). *New Zealand Journal of Geology and Geophysics*.  
<https://doi.org/10.1080/00288306.2021.1960865>
- Caron V, Bailleul J, Chanier F, Mahieux G, Joanny F-X. 2019. A new analytical procedure to graphically characterize the taphonomic properties of skeletal carbonates. An example from Miocene limestones of new zealand. *PALAIOS*. 34(8):364–381.  
<https://doi.org/10.2110/palo.2018.101>
- Caron V, Nelson CS. 2009. Diversity of Neomorphic Fabrics in New Zealand Pliocene Pleistocene Cool-Water Limestones: Insights into Aragonite Alteration Pathways and Controls. *Journal of Sedimentary Research*. 79(4):226–246.  
<https://doi.org/10.2110/jsr.2009.029>
- Caron V, Nelson CS, Kamp PJJ. 2004. Contrasting carbonate depositional systems for Pliocene cool-water limestones cropping out in central Hawke's Bay, New Zealand. *New Zealand Journal of Geology and Geophysics*. 47(4):697–717.  
<https://doi.org/10.1080/00288306.2004.9515084>
- Chanier F. 1991. Le prisme d'accr tion Hikurangi: un t moin de l' volution g odynamique d'une marge active pacifique (Nouvelle-Z lande) [PhD Thesis]. France: Universit  de Lille 1.

- Chanier F, Bellier J-P, Bignot G, Ferrière J. 1990. Discovery of Pithonellids in the Upper Cretaceous of New Zealand; biogeographic, stratigraphic and tectonic consequences. *Comptes Rendus de l'Académie des Sciences - Séries II - Earth and Planetary Science*. 310(2):1095–1100.
- Chanier F, Ferrière J. 1991. From a passive to an active margin: Tectonic and sedimentary processes linked to the birth of an accretionary prism (Hikurangi Margin, New Zealand). *Société Géologique de France*. 162(4):649–660. <https://doi.org/10.2113/gssgfbull.162.4.649>
- Chanier F, Ferrière J, Angelier J. 1999. Extensional deformation across an active margin, relations with subsidence, uplift, and rotations: The Hikurangi subduction, New Zealand. *Tectonics*. 18(5):862–876. <https://doi.org/10.1029/1999TC900028>
- Chiarella D, Longhitano SG. 2012. Distinguishing Depositional Environments In Shallow-Water Mixed, Bio-Siliciclastic Deposits On the Basis Of The Degree Of Heterolithic Segregation (Gelasian, Southern Italy) DEGREE OF HETEROLITHIC SEGREGATION IN MIXED DEPOSITS. *Journal of Sedimentary Research*. 82(12):969–990. <https://doi.org/10.2110/jsr.2012.78>
- Chiarella D, Longhitano SG, Sabato L, Tropeano M. 2012. Sedimentology and hydrodynamics of mixed (siliciclastic-bioclastic) shallow-marine deposits of Acerenza (Pliocene, Southern Apennines, Italy). *Italian Journal of Geosciences*. 131(1):136–151. <https://doi.org/10.3301/IJG.2011.36>
- Chiarella D, Longhitano SG, Tropeano M. 2017. Types of mixing and heterogeneities in siliciclastic-carbonate sediments. *Marine and Petroleum Geology*. 88:617–627. <https://doi.org/10.1016/j.marpetgeo.2017.09.010>
- Chiarella D, Longhitano SG, Tropeano M. 2019. Different stacking patterns along an active fold-and-thrust belt—Acerenza Bay, Southern Apennines (Italy). *Geology*. 47(2):139–142. <https://doi.org/10.1130/G45628.1>
- Claussmann B, Bailleul J, Chanier F, Mahieux G, Caron V, McArthur AD, Chaptal C, Morgans HEG, Vendeville BC. 2021. Shelf-derived mass-transport deposits: origin and significance in the stratigraphic development of trench-slope basins. *New Zealand Journal of Geology and Geophysics*. 1:1–36. <https://doi.org/10.1080/00288306.2021.1918729>
- Ćosović V, Mrinjek E, Nemeč W, Španiček J, Terzić K. 2018. Development of transient carbonate ramps in an evolving foreland basin. *Basin Research*. 30(4):746–765. <https://doi.org/10.1111/bre.12274>
- Crisóstomo-Figueroa A, McArthur AD, Dorrell RM, Amy L, McCaffrey WD. 2020. A new modelling approach to sediment bypass prediction applied to the East Coast Basin, New Zealand. *GSA Bulletin*. <https://doi.org/10.1130/B35687.1>
- Crundwell M. 1987. Neogene stratigraphy and geological history of the Wainuioru Valley, Eastern Wairarapa, New Zealand. [BSc Thesis]. Wellington, New Zealand: Victoria University.
- Cumberpatch ZA, Soutter EL, Kane IA, Casson M, Vincent SJ. 2021. Evolution of a mixed siliciclastic-carbonate deep-marine system on an unstable margin: The Cretaceous of the Eastern Greater Caucasus, Azerbaijan. *Basin Research*. 33(1):612–647. <https://doi.org/10.1111/bre.12488>
- Daly RA. 1936. Origin of submarine canyons. *American Journal of Science*. s5-31(186):401–420. <https://doi.org/10.2475/ajs.s5-31.186.401>
- Davies DJ, Powell EN, Stanton RJ. 1989. Relative rates of shell dissolution and net sediment accumulation - a commentary: can shell beds form by the gradual accumulation of biogenic debris on the sea floor? *Lethaia*. 22(2):207–212. <https://doi.org/10.1111/j.1502-3931.1989.tb01683.x>

- Deptuck ME, Steffens GS, Barton M, Pirmez C. 2003. Architecture and evolution of upper fan channel-belts on the Niger Delta slope and in the Arabian Sea. *Marine and Petroleum Geology*. 20(6):649–676.  
<https://doi.org/10.1016/j.marpetgeo.2003.01.004>
- Di Celma C, Cantalamessa G, Didaskalou P, Lori P. 2010. Sedimentology, architecture, and sequence stratigraphy of coarse-grained, submarine canyon fills from the Pleistocene (Gelasian-Calabrian) of the Peri-Adriatic basin, central Italy. *Marine and Petroleum Geology*. 27(7):1340–1365.  
<https://doi.org/10.1016/j.marpetgeo.2010.05.011>
- Festa A, Pini GA, Dilek Y, Codegone G. 2010. Mélanges and mélange-forming processes: a historical overview and new concepts. *International Geology Review*. 52(10–12):1040–1105.  
<https://doi.org/10.1080/00206810903557704>
- Field B, Uruski CI, Institute of Geological & Nuclear Sciences. 1997. Cretaceous-Cenozoic geology and petroleum systems of the East Coast region, New Zealand. New Zealand.
- Fisher RV. 1983. Flow transformations in sediment gravity flows. *Geology*. 11(5):273–274.
- Fonnesu M, Patacci M, Haughton PDW, Felletti F, McCaffrey WD. 2016. Hybrid Event Beds Generated By Local Substrate Delamination On A Confined-Basin Floor. *Journal of Sedimentary Research*. 86(8):929–943.  
<https://doi.org/10.2110/jsr.2016.58>
- Galloway WE. 1998. Siliciclastic Slope and Base-of-Slope Depositional Systems: Component Facies, Stratigraphic Architecture, and Classification. *AAPG Bulletin*. 824:569–595.  
<https://doi.org/10.1306/1D9BC5BB-172D-11D7-8645000102C1865D>
- Gamberi F, Rovere M, Dykstra M, Kane IA, Kneller BC. 2013. Integrating modern seafloor and outcrop data in the analysis of slope channel architecture and fill. *Marine and Petroleum Geology*. 41:83–103.  
<https://doi.org/10.1016/j.marpetgeo.2012.04.002>
- Gong C, Steel RJ, Wang Y, Lin C, Olariu C. 2016. Shelf-margin architecture variability and its role in sediment-budget partitioning into deep-water areas. *Earth-Science Reviews*. 154:72–101.  
<https://doi.org/10.1016/j.earscirev.2015.12.003>
- Haak AB, Schlager W. 1989. Compositional variations in calciturbidites due to sea-level fluctuations, late Quaternary, Bahamas. *Geol Rundsch*. 78(2):477–486.  
<https://doi.org/10.1007/BF01776186>
- Harris PT, Whiteway T. 2011. Global distribution of large submarine canyons: Geomorphic differences between active and passive continental margins. *Marine Geology*. 285(1):69–86.  
<https://doi.org/10.1016/j.margeo.2011.05.008>
- Hayton S, Nelson CS, Hood SD. 1995. A skeletal assemblage classification system for non-tropical carbonate deposits based on New Zealand Cenozoic limestones. *Sedimentary Geology*. 100(1):123–141.  
[https://doi.org/10.1016/0037-0738\(95\)00071-2](https://doi.org/10.1016/0037-0738(95)00071-2)
- Haywick DW. 2004. Diagenesis of polymineralic temperate limestones in a cyclothem sedimentary succession, eastern North Island, New Zealand. *New Zealand Journal of Geology and Geophysics*. 47(4):839–855.  
<https://doi.org/10.1080/00288306.2004.9515092>
- Hein FJ, Gorsline DS. 1981. Geotechnical aspects of fine-grained mass flow deposits: California Continental Borderland. *Geo-Marine Letters*. 1(1):1–5.  
<https://doi.org/10.1007/BF02463293>
- Hubbard SM, Covault JA, Fildani A, Romans BW. 2014. Sediment transfer and deposition in slope channels: Deciphering the record of enigmatic deep-sea processes from outcrop. *GSA Bulletin*. 126(5–6):857–871.



<https://doi.org/10.1130/B30996.1>

Iverson RM. 1997. The physics of debris flows. *Reviews of Geophysics*. 35(3):245–296. <https://doi.org/10.1029/97RG00426>

Janocko J, Basilici G. 2021. Architecture of coarse-grained gravity flow deposits in a structurally confined submarine canyon (late Eocene Tokaren Conglomerate, Slovakia). *Sedimentary Geology*. 417:105880. <https://doi.org/10.1016/j.sedgeo.2021.105880>

Janocko M, Nemeč W, Henriksen S, Warchoł M. 2013. The diversity of deep-water sinuous channel belts and slope valley-fill complexes. *Marine and Petroleum Geology*. 41:7–34. <https://doi.org/10.1016/j.marpetgeo.2012.06.012>

Jiao R, Seward D, Little TA, Kohn BP. 2014. Thermal history and exhumation of basement rocks from Mesozoic to Cenozoic subduction cycles, central North Island, New Zealand. *Tectonics*. 33(10):1920–1935. <https://doi.org/10.1002/2014TC003653>

Johnson AM. 1984. Debris flow. In: Brunnsden D, Prior DB, editors. *Slope Instability*. New York, America: Wiley and Sons; p. 257–361.

Johnston MR. 1980. *Geology of the Tinui-Awatoitoi district*. Wellington, New Zealand: New Zealand Dept of Scientific and Industrial Research.

Kneller B. 1995. Beyond the turbidite paradigm: physical models for deposition of turbidites and their implications for reservoir prediction. *Geological Society, London, Special Publications*. 94(1):31–49. <https://doi.org/10.1144/GSL.SP.1995.094.01.04>

Kneller B, Bozetti G, Callow R, Dykstra M, Hansen L, Kane I, Li P, McArthur A, Catharina AS, Santos TD, Thompson P. 2020. Architecture, process, and environmental diversity in a late cretaceous slope channel system. *JSedimentRes*. 90(1):1–26. <https://doi.org/10.2110/jsr.2020.1>

Kowalewski M. 1996. Time-Averaging, Overcompleteness, and the Geological Record. *The Journal of Geology*. 104(3):317–326. <https://doi.org/10.1086/629827>

Kuenen PhH. 1964. Deep-Sea Sands and Ancient Turbidites. In: Bouma AH, Brouwer A, editors. *Developments in Sedimentology*. Vol. 3. Netherlands: Elsevier; p. 3–33. [https://doi.org/10.1016/S0070-4571\(08\)70953-1](https://doi.org/10.1016/S0070-4571(08)70953-1)

Lackey JK, Regalla CA, Moore GF. 2020. Tectonic Influences on Trench Slope Basin Development via Structural Restoration Along the Outer Nankai Accretionary Prism, Southwest Japan. *Geochemistry, Geophysics, Geosystems*. 21(8):1–17.

Lee J, Begg J. 2002. *Geology of the Wairarapa area*. Institute of Geological & Nuclear Sciences 1:250 000 geological map: Institute of Geological & Nuclear Sciences Limited.

Lewis KB, Pettinga JR. 1993. The emerging, imbricate frontal wedge of the Hikurangi margin. In: Ballance PF, editor. *South Pacific sedimentary basins*. Amsterdam, Netherlands: Elsevier Science; p. 225–250.

Li P, Kneller B, Thompson P, Bozetti G, dos Santos T. 2018. Architectural and facies organisation of slope channel fills: Upper Cretaceous Rosario Formation, Baja California, Mexico. *Marine and Petroleum Geology*. 92:632–649. <https://doi.org/10.1016/j.marpetgeo.2017.11.026>

Lowe DR. 1982. Sediment gravity flows; II, Depositional models with special reference to the deposits of high-density turbidity currents. *Journal of Sedimentary Research*. 52(1):279–297. <https://doi.org/10.1306/212F7F31-2B24-11D7-8648000102C1865D>

Malie P, Bailleul J, Chanier F, Toullec R, Mahieux G, Caron V, Field B, Mählmann RF, Potel S. 2017. Spatial distribution and tectonic framework of fossil tubular concretions as onshore analogues of cold seep plumbing systems, North Island of New Zealand. *Bulletin*

de la Société géologique de France. 188(4):25.  
<https://doi.org/10.1051/bsgf/2017192>

Maltman A, editor. 1994. *The Geological Deformation of Sediments*. Netherlands: Springer Netherlands.  
<https://doi.org/10.1007/978-94-011-0731-0>

McArthur AD, Bailleul J, Mahieux G, Clausmann B, Wunderlich A, McCaffrey WD. 2021. Deformation-sedimentation feedback and the development of anomalously thick aggradational turbidite lobes: subsurface and outcrop examples from the Hikurangi Margin, New Zealand. *Journal of Sedimentary Research*. 91(4):362–389.  
<https://doi.org/10.2110/jsr.2020.013>

McArthur AD, Clausmann B, Bailleul J, McCaffrey W, Clare A. 2019. Variation in syn-subduction sedimentation patterns from inner to outer portions of deep-water fold and thrust belts: examples from the Hikurangi subduction margin of New Zealand. *Geological Society, London, Special Publications*. 490:285–310.  
<https://doi.org/10.1144/SP490-2018-95>

McArthur Adam D., Jolley DW, Hartley AJ, Archer SG, Lawrence HM. 2016a. Palaeoecology of syn-rift topography: A Late Jurassic footwall island on the Josephine Ridge, Central Graben, North Sea. *Palaeogeogr Palaeoclimatol Palaeoecol*. 459:63–75.  
<https://doi.org/10.1016/j.palaeo.2016.06.033>

McArthur A. D., Kneller BC, Souza PA, Kuchle J. 2016b. Characterization of deep-marine channel-levee complex architecture with palynofacies: An outcrop example from the Rosario Formation, Baja California, Mexico. *Marine and Petroleum Geology*. 73:157–173.  
<https://doi.org/10.1016/j.marpetgeo.2016.02.030>

McArthur AD, McCaffrey WD. 2019. Sedimentary architecture of detached deep-marine canyons: Examples from the East Coast Basin of New Zealand. *Sedimentology*. 66(3):1067–1101.  
<https://doi.org/10.1111/sed.12536>

McHargue T, Pycz MJ, Sullivan MD, Clark JD, Fildani A, Romans BW, Covault JA, Levy M, Posamentier HW, Drinkwater NJ. 2011. Architecture of turbidite channel systems on the continental slope: Patterns and predictions. *Marine and Petroleum Geology*. 28(3):728–743.  
<https://doi.org/10.1016/j.marpetgeo.2010.07.008>

Mohrig D, Ellis C, Parker G, Whipple KX, Hondzo M. 1998. Hydroplaning of subaqueous debris flows. *GSA Bulletin*. 110(3):387–394.  
[https://doi.org/10.1130/0016-7606\(1998\)110<0387:HOSDF>2.3.CO;2](https://doi.org/10.1130/0016-7606(1998)110<0387:HOSDF>2.3.CO;2)

Mohrig D, Marr JD. 2003. Constraining the efficiency of turbidity current generation from submarine debris flows and slides using laboratory experiments. *Marine & Petroleum Ecology*. 20(6–8):883–899.  
<https://doi.org/10.1016/j.marpetgeo.2003.03.002>

Moore GF, Karig DE. 1976. Development of sedimentary basins on the lower trench slope. *Geology*. 4(11):693–697.  
[https://doi.org/10.1130/0091-7613\(1976\)4<693:DOSBOT>2.0.CO;2](https://doi.org/10.1130/0091-7613(1976)4<693:DOSBOT>2.0.CO;2)

Moscardelli L, Ochoa J, Lunt I, Zahm L. 2019. Mixed siliciclastic–carbonate systems and their impact for the development of deep-water turbidites in continental margins: A case study from the Late Jurassic to Early Cretaceous Shelburne subbasin in offshore Nova Scotia. *AAPG Bulletin*. 103(10):2487–2520.  
<https://doi.org/10.1306/02151917318>

Moscardelli L, Wood L. 2008. New classification system for mass transport complexes in offshore Trinidad. *Basin Research*. 20(1):73–98.  
<https://doi.org/10.1111/j.1365-2117.2007.00340.x>

Moscardelli L, Wood L. 2015. Morphometry of mass-transport deposits as a predictive tool. *GSA Bulletin*. 128(1–2):47–80.

Mountjoy JJ, Barnes PM, Pettinga JR. 2009. Morphostructure and evolution of submarine canyons across an active margin: Cook Strait

sector of the Hikurangi Margin, New Zealand. *Marine Geology*. 260(1):45–68. <https://doi.org/10.1016/j.margeo.2009.01.006>

Mountjoy JJ, Howarth JD, Orpin AR, Barnes PM, Bowden DA, Rowden AA, Schimel ACG, Holden C, Horgan HJ, Nodder SD, et al. 2018. Earthquakes drive large-scale submarine canyon development and sediment supply to deep-ocean basins. *Science Advances*. 4(3):1–9. <https://doi.org/10.1126/sciadv.aar3748>

Mulder T, Alexander J. 2001. The physical character of subaqueous sedimentary density flows and their deposits. *Sedimentology*. 48(2):269–299. <https://doi.org/10.1046/j.1365-3091.2001.00360.x>

Mulder T, Cochonat P. 1996. Classification of offshore mass movements. *Journal of Sedimentary Research*. 66(1):43–57. <https://doi.org/10.1306/D42682AC-2B26-11D7-8648000102C1865D>

Mulder T, Ducassou E, Gillet H, Hanquiez V, Tournadour E, Combes J, Eberli GP, Kindler P, Gonthier E, Conesa G, et al. 2012. Canyon morphology on a modern carbonate slope of the Bahamas: Evidence of regional tectonic tilting. *Geology*. 40(9):771–774. <https://doi.org/10.1130/G33327.1>

Mulder T, Joumes M, Hanquiez V, Gillet H, Reijmer JJG, Tournadour E, Chabaud L, Principaud M, Schnyder JSD, Borgomano J, et al. 2017. Carbonate slope morphology revealing sediment transfer from bank-to-slope (Little Bahama Bank, Bahamas). *MAR PETROL GEOL*. 83:26–34. <https://doi.org/10.1016/j.marpetgeo.2017.03.002>

Mutti E. 1985. Turbidite Systems and Their Relations to Depositional Sequences. In: Zuffa GG, editor. *Provenance of Arenites*. Dordrecht, Netherlands: Springer Netherlands; p. 65–93. [https://doi.org/10.1007/978-94-017-2809-6\\_4](https://doi.org/10.1007/978-94-017-2809-6_4)

Naranjo-Vesga J, Ortiz-Karpf A, Wood L, Jobe Z, Paniagua-Aroyave JF, Shumaker L, Mateus-Tarazona D, Galindo P. 2020. Regional controls

in the distribution and morphometry of deep-water gravitational deposits along a convergent tectonic margin. *Southern Caribbean of Colombia. Marine and Petroleum Geology*. 121:104639.

<https://doi.org/10.1016/j.marpetgeo.2020.104639>

Nardin TR, Hein FJ, Gorsline DS, Edwards BD. 1979. A review of mass movement processes, sediment and acoustic characteristics, and contrasts in slope and base-of-slope systems versus canyon-fan-basin floor systems. *SEPM Special Publication*. 27:61–73.

Nebelsick JH, Rasser MW, Bassi D. 2005. Facies dynamics in Eocene to Oligocene circumalpine carbonates. *Facies*. 51(1):197–217. <https://doi.org/10.1007/s10347-005-0069-2>

Neef G. 1992. Geology of the Akitio area (1:50 000 metric sheet U25BD, east), northeastern Wairarapa, New Zealand. *New Zealand Journal of Geology and Geophysics*. 35(4):533–548. <https://doi.org/10.1080/00288306.1992.9514546>

Neef G. 1999. Neogene development of the onland part of the forearc in northern Wairarapa, North Island, New Zealand: A synthesis. *New Zealand Journal of Geology and Geophysics*. 42(1):113–135. <https://doi.org/10.1080/00288306.1999.9514835>

Nemec W, Steel RJ. 1984. Alluvial and Coastal Conglomerates: Their Significant Features and Some Comments on Gravelly Mass-Flow Deposits. In: Koster EH, Steel RJ, editors. *Sedimentology of Gravels and Conglomerates*, Memoir 10. Canada: Canadian Society of Petroleum Geologists; p. 1–31.

Nicol A, Mazengarb C, Chanier F, Rait G, Uruski C, Wallace L. 2007. Tectonic evolution of the active Hikurangi subduction margin, New Zealand, since the Oligocene. *Tectonics*. 26(4):1–24.

<https://doi.org/10.1029/2006TC002090>



- Nicol A, VanDissen R, Vella P, Alloway B, Melhuish A. 2002. Growth of contractional structures during the last 10 m.y. at the southern end of the emergent Hikurangi forearc basin, New Zealand. *New Zealand Journal of Geology and Geophysics*. 45(3):365–385. <https://doi.org/10.1080/00288306.2002.9514979>
- Nicolaidis S. 1995. Cementation in Oligo-Miocene non-tropical shelf limestones, Otway Basin, Australia. *Sedimentary Geology*. 95:97–121. [https://doi.org/10.1016/0037-0738\(94\)00102-Z](https://doi.org/10.1016/0037-0738(94)00102-Z)
- Ogata K, Festa A, Pini GA, Pogačnik Ž, Lucente CC. 2019. Substrate deformation and incorporation in sedimentary mélanges (olistostromes): Examples from the northern Apennines (Italy) and northwestern Dinarides (Slovenia). *Gondwana Research*. 74:101–125. <https://doi.org/10.1016/j.gr.2019.03.001>
- Ortiz-Karpf A, Hodgson DM, Jackson CA-L, McCaffrey WD. 2017. Influence of Seabed Morphology and Substrate Composition On Mass-Transport Flow Processes and Pathways: Insights From the Magdalena Fan, Offshore Colombia. *Journal of Sedimentary Research*. 87(3):189–209. <https://doi.org/10.2110/jsr.2017.10>
- Ortiz-Karpf A, Hodgson DM, Jackson CA-L, McCaffrey WD. 2018. Mass-transport complexes as markers of deep-water fold-and-thrust belt evolution: insights from the southern Magdalena fan, offshore Colombia. *Basin Research*. 30(S1):65–88. <https://doi.org/10.1111/bre.12208>
- Paumard V, Bourget J, Payenberg T, George AD, Ainsworth RB, Lang S, Posamentier HW. 2020. Controls On Deep-Water Sand Delivery Beyond the Shelf Edge: Accommodation, Sediment Supply, and Deltaic Process Regime. *Journal of Sedimentary Research*. 90(1):104–130. <https://doi.org/10.2110/jsr.2020.2>
- Peakall J, Sumner EJ. 2015. Submarine channel flow processes and deposits: A process-product perspective. *Geomorphology*. 244:95–120. <https://doi.org/10.1016/j.geomorph.2015.03.005>
- Perry CT. 1998. Grain susceptibility to the effects of microboring: implications for the preservation of skeletal carbonates. *Sedimentology*. 45(1):39–51. <https://doi.org/10.1046/j.1365-3091.1998.00134.x>
- Pettinga JR. 1982. Upper Cenozoic structural history, coastal Southern Hawke's Bay, New Zealand. *New Zealand Journal of Geology and Geophysics*. 25(2):149–191. <https://doi.org/10.1080/00288306.1982.10421407>
- Pirmez C, Beaubouef RT, Friedmann SJ, Mohrig DC. 2000. Equilibrium Profile and Baselevel in Submarine Channels: Examples from Late Pleistocene Systems and Implications for the Architecture of Deepwater Reservoirs. In: *Deep-Water Reservoirs of the World*. Vol. 20. Oklahoma, America: SEPM (Society for Sedimentary Geology); p. 792–805.
- Pomar L. 2001. Ecological control of sedimentary accommodation: evolution from a carbonate ramp to rimmed shelf, Upper Miocene, Balearic Islands. *Palaeogeography, Palaeoclimatology, Palaeoecology*. 175(1):249–272. [https://doi.org/10.1016/S0031-0182\(01\)00375-3](https://doi.org/10.1016/S0031-0182(01)00375-3)
- Posamentier HW, Kolla V. 2003. Seismic geomorphology and stratigraphy of depositional elements in deep-water settings. *Journal of Sedimentary Research*. 73(3):367–388.
- Posamentier HW, Martinsen OJ. 2011. The Character and Genesis of Submarine Mass-Transport Deposits: Insights from Outcrop and 3D Seismic Data. In: Shipp RC, Weimer P, Posamentier HW, editors. *Mass-Transport Deposits in Deepwater Settings*. Oklahoma, America: SEPM; p. 7–38.
- Posamentier HW, Vail PR. 1988. Eustatic Controls on Clastic Deposition II—Sequence and Systems Tract Models. In: *Sea-Level Changes: An Integrated Approach*. Vol. 42.

Oklahoma, America: SEPM (Society for Sedimentary Geology); p. 125–154.

Posamentier HW, Walker RG. 2006. Deep-Water Turbidites and Submarine Fans. In: Posamentier HW, Walker RG, editors. *Facies Models Revisited*. Oklahoma, America: SEPM Special Publication 84; p. 397–520. <https://doi.org/10.2110/pec.06.84>

Postma G. 1986. Classification for sediment gravity-flow deposits based on flow conditions during sedimentation. *Geology*. 14:291. [https://doi.org/10.1130/0091-7613\(1986\)14<291:CFSGDB>2.0.CO;2](https://doi.org/10.1130/0091-7613(1986)14<291:CFSGDB>2.0.CO;2)

Postma G, Nemeč W, Kleinspehn KL. 1988. Large floating clasts in turbidites: a mechanism for their emplacement. *Sedimentary Geology*. 58(1):47–61. [https://doi.org/10.1016/0037-0738\(88\)90005-X](https://doi.org/10.1016/0037-0738(88)90005-X)

Powers MC. 1953. A new roundness scale for sedimentary particles. *Journal of Sedimentary Research*. 23(2):117–119. <https://doi.org/10.1306/D4269567-2B26-11D7-8648000102C1865D>

Prather BE. 2003. Controls on reservoir distribution, architecture and stratigraphic trapping in slope settings. *Marine and Petroleum Geology*. 20(6):529–545. <https://doi.org/10.1016/j.marpetgeo.2003.03.009>

Prélat A, Hodgson DM, Flint SS. 2009. Evolution, architecture and hierarchy of distributary deep-water deposits: a high-resolution outcrop investigation from the Permian Karoo Basin, South Africa. *Sedimentology*. 56(7):2132–2154. <https://doi.org/10.1111/j.1365-3091.2009.01073.x>

Puigdefàbregas C, Muñoz JA, Marzo M. 1986. Thrust Belt Development in the Eastern Pyrenees and Related Depositional Sequences in the Southern Foreland Basin. In: *Foreland Basins*. United Kingdom: John Wiley & Sons, Ltd; p. 229–246.

Raine JI, Beu A, Boyes A, Campbell H, Cooper R, Crampton J, Crundwell M, Hollis C, Morgans H, Mortimer N. 2015. New Zealand Geological Timescale NZGT 2015/1. *New Zealand Journal of Geology and Geophysics*. 58(4):398–403. <https://doi.org/10.1080/00288306.2015.1086391>

Reijmer JJG, Palmieri P, Groen R, Floquet M. 2015. Calciturbidites and calcidebrites: Sea-level variations or tectonic processes? *Sedimentary Geology*. 317:53–70. <https://doi.org/10.1016/j.sedgeo.2014.10.013>

Richards M, Bowman M, Reading H. 1998. Submarine-fan systems i: characterization and stratigraphic prediction. *Marine and Petroleum Geology*. 15(7):689–717. [https://doi.org/10.1016/S0264-8172\(98\)00036-1](https://doi.org/10.1016/S0264-8172(98)00036-1)

Romero-Otero GA, Slatt RM, Pirmez C. 2010. Detached and Shelf-Attached Mass Transport Complexes on the Magdalena Deepwater Fan. In: Mosher DC, Shipp RC, Moscardelli L, Chaytor JD, Baxter CDP, Lee HJ, Urgeles R, editors. *Submarine Mass Movements and Their Consequences*. Dordrecht, Netherlands: Springer Netherlands; p. 593–606. [https://doi.org/10.1007/978-90-481-3071-9\\_48](https://doi.org/10.1007/978-90-481-3071-9_48)

Sobiesiak MS, Kneller B, Alsop GI, Milana JP. 2018. Styles of basal interaction beneath mass transport deposits. *Marine and Petroleum Geology*. 98:629–639. <https://doi.org/10.1016/j.marpetgeo.2018.08.028>

Sohn YK, Choe MY, Jo HR. 2002. Transition from debris flow to hyperconcentrated flow in a submarine channel (the Cretaceous Cerro Toro Formation, southern Chile). *Terra Nova*. 14(5):405–415.

Sømme TO, Helland-Hansen W, Martinsen OJ, Thurmond JB. 2009. Relationships between morphological and sedimentological parameters in source-to-sink systems: a basis for predicting semi-quantitative characteristics in subsurface systems. *Basin Research*. 21(4):361–387. <https://doi.org/10.1111/j.1365-2117.2009.00397.x>

Spörli KB. 1980. New Zealand and Oblique-Slip Margins: Tectonic Development up to and during the Cainozoic. In: Ballance PF, Reading HG, editors. *Sedimentation in Oblique-Slip Mobile Zones*. United Kingdom: John Wiley & Sons, Ltd; p. 147–170.

<https://doi.org/10.1002/9781444303735.ch9>

Sprague AR, Garfield TR, Goulding FJ, Beaubouef RT. 2005. Integrated slope channel depositional models: the key to successful prediction of reservoir presence and quality in Offshore West Africa. In: *Proceedings of the CIPM conference*. Vol. 5. Veracruz, Mexico; p. 1–13.

Stanley DJ. 1986. Turbidity current transport of organic-rich sediments: Alpine and Mediterranean examples. *Marine Geology*. 70(1):85–101. [https://doi.org/10.1016/0025-3227\(86\)90090-3](https://doi.org/10.1016/0025-3227(86)90090-3)

Stevens SH, Moore GF. 1985. Deformational and sedimentary processes in trench slope basins of the western Sunda Arc, Indonesia. *Marine Geology*. 69(1):93–112.

[https://doi.org/10.1016/0025-3227\(85\)90135-5](https://doi.org/10.1016/0025-3227(85)90135-5)

Strachan LJ. 2008. Flow transformations in slumps: a case study from the Waitemata Basin, New Zealand. *Sedimentology*. 55(5):1311–1332.

<https://doi.org/10.1111/j.1365-3091.2007.00947.x>

Sylvester Z, Pirmez C, Cantelli A. 2011. A model of submarine channel-levee evolution based on

channel trajectories: Implications for stratigraphic architecture. *Marine and Petroleum Geology*. 28(3):716–727.

<https://doi.org/10.1016/j.marpetgeo.2010.05.012>

Talling PJ, Amy LA, Wynn RB, Peakall J, Robinson M. 2004. Beds comprising debrite sandwiched within co-genetic turbidite: origin and widespread occurrence in distal depositional environments. *Sedimentology*. 51(1):163–194. <https://doi.org/10.1111/j.1365-3091.2004.00617.x>

Van den Heuvel HB. 1960. The geology of the flat point area, eastern Wairarapa. *New Zealand Journal of Geology and Geophysics*. 3(2):309–320.

<https://doi.org/10.1080/00288306.1960.10423603>

Watson SJ, Mountjoy JJ, Crutchley GJ. 2020. Tectonic and geomorphic controls on the distribution of submarine landslides across active and passive margins, eastern New Zealand. *Geological Society, London, Special Publications*. 500:477–494.

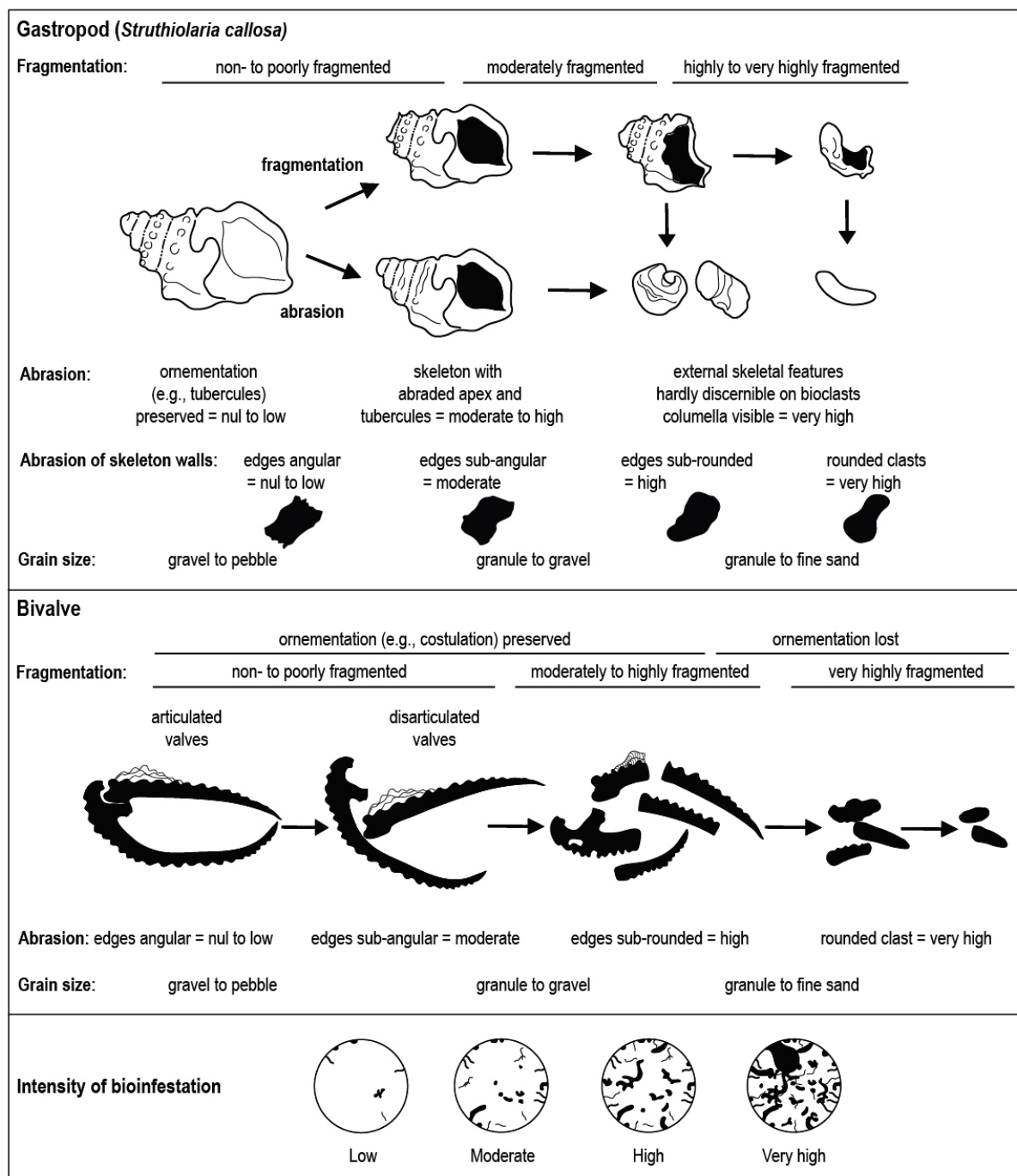
<https://doi.org/10.1144/SP500-2019-165>

Zuschin M, Stachowitsch M, Stanton RJ. 2003. Patterns and processes of shell fragmentation in modern and ancient marine environments. *Earth-Science Reviews*. 63(1):33–82.

[https://doi.org/10.1016/S0012-8252\(03\)00014-X](https://doi.org/10.1016/S0012-8252(03)00014-X)



## 11. SUPPLEMENTARY MATERIAL



Appendix 1: Taphonomic methodology used on the macrofossils observed in the Waikaraka and Homewood localities. Qualitative taphonomic grading system established for the two dominant categories of fossil remains found in the studied deposits, namely Gastropods and Bivalves. The scales of alteration features were established to help characterize poorly to very highly damaged grains at outcrop. The particle contours displayed derive from field observations. The degree of abrasion was determined using the visual chart of (Powers 1953). The scale of skeleton bio-infestation is modified from (Perry 1998). The photographs provide field examples.

Sections	Fossil record Nb. & Field Nb.	LF code	Collector	Identifier	Date	Age	Age comment	Paleoecology & Paleobathymetry	Longitude	Latitude
	T27/0631 18_05_01	MDST	Clausmann, B.	Morgans, H.E.G. Gard, H.J.L.	02/2018	Upper Sc Lower Si	Planktic fauna suggests Late Cretaceous to Early Lillburnian (Middle Miocene; ca. 15.5 – ca. 14 Ma). No age diagnostic molluscs were present. Stratigraphically located above sample 18_05_06 and below T27/0493.	Deep middle bathyal (>800 m) or deeper. Planktic abundance >90% indicates a fully oceanic water mass overlay the site of deposition.	175,99163300	-41,19168700
<b>Homewood-1 section</b>	T27/0634 18_05_06	DF	Clausmann, B.	Morgans, H.E.G. Gard, H.J.L.	02/2018	SI - Tt	Broadest age is Lillburnian to Early T ongaporutuan (Middle to Late Miocene; Late Langhien - Tortonian; 15.1 - 7.2 Ma). The absence of <i>Paraglobobolalia mayeri</i> could indicate Early T ongaporutuan. Stratigraphically located below sample 18_05_01.	Middle bathyal paleodepth (>700 m). Although not abundant, the planktic fauna is 95% of the total foraminiferal fauna and indicates a fully oceanic water mass.	175,99264500	-41,18961700
	T27/0493	DF	Crundwell, M.P. Watson, J.	Crundwell, M.P.	01/1992	SI	Restricted to Lillburnian age (Middle Miocene; Late Langhien - Serravalian; 15.1 - 13.05 Ma). Stratigraphically located above sample 18_05_01 and 18_05_06.	Upper bathyal. Planktic abundance is 20%. Marginal oceanic slightly anoxic bottom conditions. Abundant Thalmammalia-like taxa (turbidite fauna?).	175,99248°E	41,1905°S
	T27/0625 18_03_02a	DF/CF	Clausmann, B. Maheux, G.	Morgans, H.E.G. Gard, H.J.L.	02/2018	SI - Tt	No planktics were recovered. The benthics indicate an age from Runangan to Waipian (Late Eocene to Pliocene; 34.6 - 3.00 Ma). Associated molluscs ( <i>Stropholaria callosa</i> ) give a Lillburnian to T ongaporutuan age (Middle to Late Miocene; Late Langhien - Tortonian; 15.1 - 7.2 Ma).	The presence of <i>Cibicides molestus</i> indicates deep outer bathyal paleodepth (>150 m). The presence of molluscs indicates shallow waters.	176,03276700	-41,15014600
	T27/0626 18_03_02b	DF/CF	Clausmann, B. Maheux, G.	Morgans, H.E.G. Gard, H.J.L.	02/2018	Lw - Sw	Only benthic recovered has a range of Waikikian to Waiauan (Late Oligocene to Middle Miocene; Late Chaitian - Serravalian; 25.2 - 11.04 Ma). No age diagnostic molluscs were present.	No real data. The molluscs indicate shallow waters and originate from mid to inner shelf environment.	176,03276700	-41,15014600
<b>Waikaraka section</b>	T27/0627 18_03_02c	DF/CF	Clausmann, B. Maheux, G.	Morgans, H.E.G. Gard, H.J.L.	02/2018	SI - Tt	A very poor foraminiferal fauna suggests Runangan to Recent (Late Eocene - Recent; 36.7 - 0 Ma). Molluscs give a Lillburnian to T ongaporutuan age (Middle to Late Miocene; Late Langhien - Tortonian; 15.1 - 7.2 Ma).	No real data. The molluscs indicate shallow waters and originate from mid to inner shelf environment.	176,03276700	-41,15014600
	T27/0628 18_03_02e	DF/CF	Clausmann, B. Maheux, G.	Morgans, H.E.G. Gard, H.J.L.	02/2018	NA	Only some possible agglutinated taxa and a nodosanid were observed.	Marine.	176,03276700	-41,15014600
	T27/0644 20_16_17	LDTc	Clausmann, B. Maheux, G.	Morgans, H.E.G.	12/2020	Sw	A good planktic fauna was recovered including a sinistrally coiled population of <i>Globobolalia mitulimida</i> , <i>Globoquadrina dehiscentis</i> and <i>Paraglobobolalia mayeri</i> . The age is Waiauan (Middle Miocene; Serravalian; 13.05 - 11.63).	The presence of <i>Tritaxilina zealandica</i> , <i>Sigmatopsis schlumbergeri</i> , <i>Karriella cylindrica</i> indicates paleodepths at 1500 metres or deeper (lower bathyal). Abundant <i>Amphisigina</i> , and less common <i>Zenithorinus</i> , <i>Florilus</i> , <i>Ehrenbergina</i> , and <i>Cibicides perforatus</i> indicates shallow to middle shelf reworking across the shelf into deeper water. Planktic abundance is about 40% indicating a middle neritic watermass.	176,02902333	-41,15342233

Appendix 2: Sample list used for the Whareama Basin outcrops of Homewood and Waikaraka. The Fossil Record numbers including "T27" relate to the T27 New Zealand Map Grid 1:50,000 series and are captured in the Fossil Record Electronic Database (<https://fred.org.nz>). See Table 1 for the description of the associated lithofacies code. The age of the sedimentary units and the paleobathymetries were defined using micro- and macropaleontological analyses conducted by GNS Science New Zealand.

**(A) Macropaleontology list**

Fossil record No.	T27/0631	T27/0634		T27/0625	T27/0626	T27/0627	
Field No.	18_05_01	18_05_06		18_03_02a	18_03_02b	18_03_02c	
Identifier	Gard, H.J.L.	Gard, H.J.L.		Gard, H.J.L.	Gard, H.J.L.	Gard, H.J.L.	
Identification data	11/2018	11/2018		11/2018	11/2018	11/2018	
Stage - Start (older)	Sc	SI	General observations throughout entire outcrop deposits	SI	Lw	SI	General observations throughout entire outcrop deposits
Stage - Start (younger)	MDST	Tt		Tt	Sw	Tt	
Lithofacies code	MDST	DF		DF/CF	DF/CF	DF/CF	
Locality	Homewood	Homewood		Waikaraka	Waikaraka	Waikaraka	
Trench-slope basin	Whareama	Whareama		Whareama	Whareama	Whareama	
Laboratory	GNS Science	GNS Science		GNS Science	GNS Science	GNS Science	
<b>Macrofauna</b>							
Molluscs (gastropod)	<i>Austrofusus</i> sp. (?)	x				x	
Molluscs (gastropod)	<i>Naticidae</i>						
Molluscs (gastropod)	<i>Policines</i> sp.		x				x
Molluscs (gastropod)	<i>Struthiolaria (Callusaria) callosa</i>		x	common		x	x
Molluscs (gastropod)	<i>Struthiolaria aff. spinosa</i>				x		
Molluscs (gastropod)	<i>Struthiolaria</i> sp.		x				x
Molluscs (gastropod)	<i>Tumitella</i> sp.	x	x				x
Molluscs (gastropod)	<i>Volutidae</i>				x		
Molluscs (bivalvia)	<i>Cardium</i> sp.		x				x
Molluscs (bivalvia)	<i>Glycymerididae</i>			x	x		
Molluscs (bivalvia)	<i>Glycymeris</i> sp.		x	x			x
Molluscs (bivalvia)	<i>Glycymeria</i> sp. (?)	x					
Molluscs (bivalvia)	<i>Ostrea</i> sp.		x				x
Corals	<i>Oculina</i> sp. (?)	x	x				
Corals	Colonial corals	x	x				
Corals	Solitary corals						x
Marine mammals	<i>Undifferentiated (bones)</i>		x				
Rhodolith	/		x				x

**(B) Micropaleontology list**

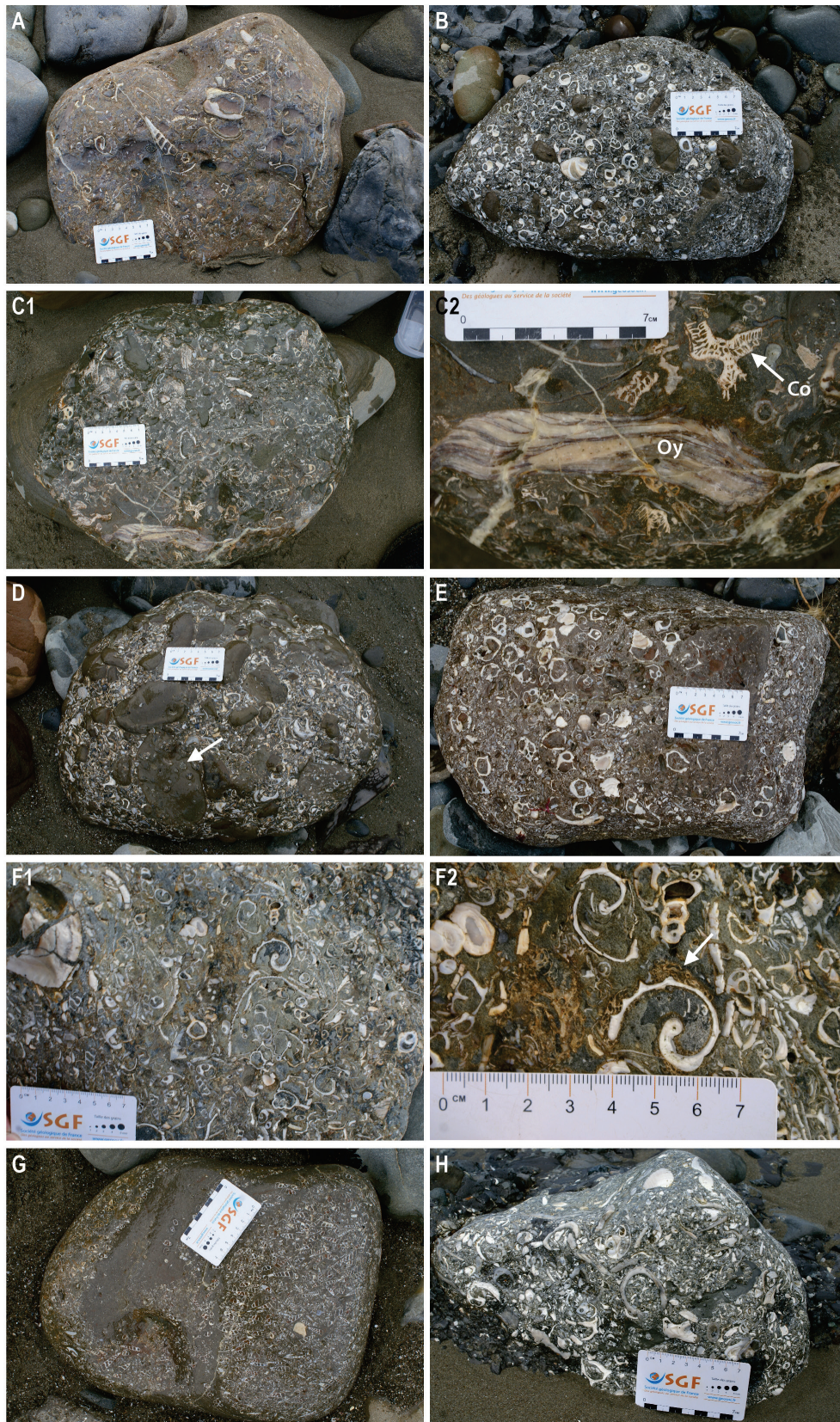
Fossil record No.	T27/0631	T27/0634	T27/0493	T27/0625	T27/0626	T27/0627	T27/0628	T27/0644
Field No.	18_05_01	18_05_06	/	18_03_02a	18_03_02b	18_03_02c	18_03_02c	20_16_17
Identifier	Morgans, H.E.G.	Morgans, H.E.G.	Crundwell, M.P.	Morgans, H.E.G.	Morgans, H.E.G.	Morgans, H.E.G.	Morgans, H.E.G.	Morgans, H.E.G.
Identification data	11/2018	11/2018	01/1992	11/2018	11/2018	11/2018	11/2018	12/2020
Stage - Start (older)	Sc	SI	SI	SI	Lw	SI	NA	Sw
Stage - Start (younger)	SI	Tt	SI	Tt	Sw	Tt	NA	Sw
Lithofacies code	MDST	DF	DF	DF/CF	DF/CF	DF/CF	DF/CF	LDTC
Locality	Homewood	Homewood	Homewood	Waikaraka	Waikaraka	Waikaraka	Waikaraka	Waikaraka
Trench-slope basin	Whareama	Whareama	Whareama	Whareama	Whareama	Whareama	Whareama	Whareama
Laboratory	GNS Science	GNS Science	GNS Science	GNS Science	GNS Science	GNS Science	GNS Science	GNS Science
<b>Microfauna</b>								
Foraminifera	<i>Amphispirina</i> sp.	x						x
Foraminifera	<i>Anomalina obovatus</i>	x	x					
Foraminifera	<i>Anomalina</i> sp.							
Foraminifera	<i>Baggina angula</i>	x						
Foraminifera	<i>Biloculina</i> sp.	x						
Foraminifera	<i>Bolivina</i> sp.							x
Foraminifera	<i>Bulinina</i> sp.							x
Foraminifera	<i>Bulinina striata</i>	x						x
Foraminifera	<i>Chilostomella ovoides</i>		x					x
Foraminifera	<i>Chilostomella</i> sp.	x						x
Foraminifera	<i>Chrysalogonium striatissimum</i>	x						
Foraminifera	<i>Cibicides molestus</i>		x	x				x
Foraminifera	<i>Cibicides novozelandicus</i>	x						x
Foraminifera	<i>Cibicides perforatus</i>			x				x
Foraminifera	<i>Cibicides truncatus</i>							reworked
Foraminifera	<i>Cibicides</i> spp.	x	x					
Foraminifera	<i>Discorthis</i> sp.			x				
Foraminifera	<i>Discorthis tenuissima</i>					x		
Foraminifera	<i>Dorothyia</i> sp.							x
Foraminifera	<i>Ehrenbergina</i> sp.							x
Foraminifera	<i>Euuvigerina miozea</i>	group						x
Foraminifera	<i>Euuvigerina</i> sp.		x					x
Foraminifera	<i>Eviovoicassidulina orientalis</i>		x					x
Foraminifera	<i>Filulus stachel</i>				x			
Foraminifera	<i>Filulus</i> sp.							x
Foraminifera	<i>Folsella peripheronda</i>	x						
Foraminifera	<i>Gibberina praebuloides</i>		x					
Foraminifera	<i>Gibberina</i> spp.		x					x
Foraminifera	<i>Gibberina woodi</i>		x					x
Foraminifera	<i>Gibbulina pacifica</i>		x					x
Foraminifera	<i>Gibbulocassidulina subglobosa</i>							x
Foraminifera	<i>Gibboquadria dehiscentis</i>		x					x
Foraminifera	<i>Gibborotalia miodumida</i>	sinistral	x					sinistral
Foraminifera	<i>Gibborotalia miozea</i>	x						
Foraminifera	<i>Gibborotalia panda</i>			x				
Foraminifera	<i>Gibborotalia praemendii</i>			x				
Foraminifera	<i>Gyrogonia zealandica</i>							x
Foraminifera	<i>Gyrogonia neosidani</i>	x	x					x
Foraminifera	<i>Haeslerella pukariensis</i>							x
Foraminifera	<i>Haplophragmoides</i> sp.					x		
Foraminifera	<i>Hoegundina elegans</i>	x	x					
Foraminifera	<i>Karreriella cylindrica</i>	x						x
Foraminifera	<i>Karreriella</i> sp.	x						
Foraminifera	<i>Laticulina pasperata</i>	x						
Foraminifera	<i>Lenticulina echinatus</i>							x
Foraminifera	<i>Lenticulina gyrocalpus</i>							x
Foraminifera	<i>Lenticulina</i> sp.	x	x					
Foraminifera	<i>Lingulina avellanoides</i>	x						
Foraminifera	<i>Marinittella communis</i>	x						
Foraminifera	<i>Melonis zeabesius</i>							x
Foraminifera	<i>Nodosaria callosa</i>							x
Foraminifera	<i>Nodosaria longiscala</i>	x						x
Foraminifera	<i>Nodosaria</i> sp.	x	x		x		x	x
Foraminifera	<i>Notorotalia taranaki</i>							x
Foraminifera	<i>Notorotalia</i> spp.							x
Foraminifera	<i>Otubulina</i> sp.		x					
Foraminifera	<i>Otubulina suturalis</i>	x		x				
Foraminifera	<i>Otubulina universa</i>							x
Foraminifera	<i>Otidorsalis umbonatus</i>			x				x
Foraminifera	<i>Ozargulatia culter</i>							x
Foraminifera	<i>Parafissurina</i> sp.	x						
Foraminifera	<i>Paragibborotalia mayeri</i>	x						x
Foraminifera	<i>Pararotalia mackayi</i>							x
Foraminifera	<i>Parvicarinina altocamerata</i>	x						
Foraminifera	<i>Pleurostomella</i> sp.		x					x
Foraminifera	<i>Pleurostomella</i> sp.	x						
Foraminifera	<i>Pulvinella buloides</i>	x	x					x
Foraminifera	<i>Quinqueloculina</i> spp.	x	x					x
Foraminifera	<i>Sigmillopsis schlumbergeri</i>	x	x					x
Foraminifera	<i>Siphovigerina plebeja</i>	x	x					x
Foraminifera	<i>Sphaeroidina bulloides</i>			x		x		x
Foraminifera	<i>Sphaeroidina dehiscentis</i>							x
Foraminifera	<i>Sibicostomella pomulgera</i>	x						
Foraminifera	<i>Taxidaria glabra</i>							x
Foraminifera	<i>Taxidaria miozea</i>							x
Foraminifera	<i>Tritaxina zealandica</i>							x
Foraminifera	<i>Trochamminoides</i> sp.	x						
Foraminifera	<i>Vaginulina</i> sp.							x
Foraminifera	<i>Vaginulinopsis</i> spp.	x						
Foraminifera	<i>Zealofus</i> sp.							x

Appendix 3: List of the macro-organisms & micro-organisms observed in each sample during the analyses conducted by GNS New Zealand. This information is captured in the Fossil Record Electronic Database (<https://fred.org.nz/>).



Appendix 4: Enlarged view of the Homewood outcrop, mostly comprising a several hundreds of meter-thick distal depositional lobe system (Fa1g). This study particularly focuses on the gravity-driven systems present at the base of this series and which are presented in the detailed view of the outcrop model in Figure 9.





Appendix 5: Clasts of the inner- to mid-shelf shell beds clasts encountered at Waikaraka. (A): Siliciclastic floatstone dominated by gastropods (*Struthiolaria* sp. and turrrellids) and thick-shelled bivalves in a sandstone matrix. (B): Pebbly molluscan rudstone. Sandstone clasts include sub-rounded to rounded gravels and pebbles. (C1-2): High-diversity clast-supported



rudstone containing thick-shelled oysters (Oy, C2), zooxanthellate branching corals (Co, C2), brachiopods, turrnellids, serpulids and subsidiary bryozoans. (D): Poorly-sorted pebbly rudstone containing abundant bored, well-rounded, sandy pebbles, and highly fragmented and abraded molluscan remains. (E): Molluscan floatstone-rudstone rich in thick-shelled gastropods and bivalves. (F1-2): Molluscan rudstone includes both thin- and thick-shelled molluscs commonly encrusted by coralline red algae (arrow, F2). (G): Turrnellid floatstone. (H): Densely packed, thick-shelled molluscan rudstone.



### Calculations using Moscardelli & Wood 2015

#### GENERAL

Equation (1):  $L = 1.5051 * (A^{0.4977})$

Equation (4):  $V = 0.0008 * (T^{2.1998})$

Equation (12):  $V = 0.0281 * (A^{1.1123})$

Rearranged Equation (12):  $A = ((V) * (1/0.0281))^{(1/1.1123)}$

#### DETACHED

Equation (2):  $Ld = 1.2477 * (Ad^{0.5415})$

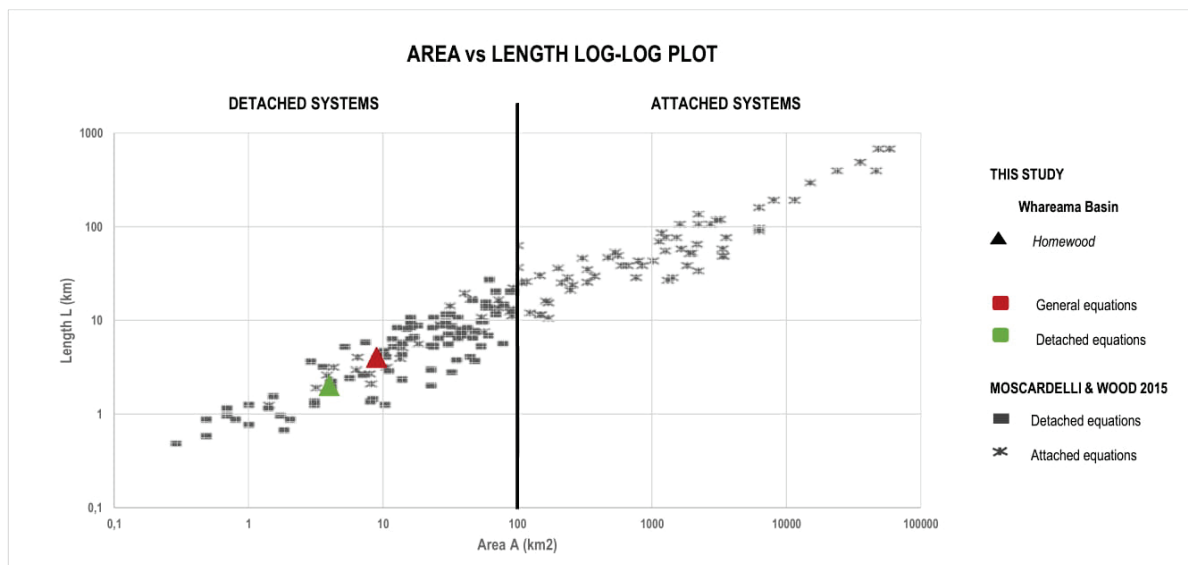
Equation (6):  $Vd = 0.0014 * (Td^{1.6223})$

Equation (14):  $Vd = 0.0261 * (Ad^{1.1507})$

Rearranged Equation (14):  $Ad = ((Vd) * (1/0.0261))^{(1/1.1507)}$

#### USING OUTCROP THICKNESS

		Thickness T (m)	Volume V (km3)	Area A (km2)	Length L (km)	Equations
WHAREAMA	INPUT	Sefton Hills-1				
		15				
						RESULT
			Equation (4)	Rearranged Equation (12)	Equation (1)	General
			0,31	9	4	
			Equation (6)	Rearranged Equation (14)	Equation (2)	Detached s.l.
			0,11	4	2	



Appendix 6: Morphometric parameter calculations using the equations from (Moscardelli and Wood 2015) for the Homewood mass-transport deposits. The outcrop conditions prevented the recognition of most of the morphometric parameters and therefore the calculations were estimated using the assumption of a relative uniform thickness of 15 meters. The calculations were done using the general set of equations as well as the dedicated set for “detached systems”.