Sedimentology (1979) **26**, 475–496

Trench-slope channels from the New Zealand Jurassic: the Otekura Formation, Sandy Bay, South Otago

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ABSTRACT

The Otekura Formation (Early Jurassic, *Pseudaucella* zone) at Sandy Bay comprises part of a 10+ km thick, regressive, forearc shelf and slope sequence, the Hokonui facies belt of the Rangitata Geosyncline. The Otekura Formation is dominantly fine grained, being mostly mudstone, silty mudstone and siltstone. The sediments are volcanogenic throughout. The upper 150 m of the formation contains two 20 m thick, channelized bodies of medium-thick bedded sandy flysch, each associated with thin bedded muddy flysch interpreted as overbank turbidites. Directional indicators within the channel sequence indicate emplacement from the south-southwest. In contrast, rare turbidites that occur below the channel sequence, within the background mudstone sediment, were emplaced from the east, i.e. at right angles to the channelized flows. The immediately overlying Omaru Formation contains more abundant macrofossils, intraclastic conglomerates, and appreciable amounts of traction-emplaced cross-bedded sand. Bioturbated calcareous siltstones with an *in situ* molluscan fauna follow (Boatlanding Formation), and are of shelf origin. The Omaru Formation is therefore interpreted as a shelf-slope break deposit, and the Otekura Formation as an upper slope facies.

Reconnaissance studies indicate that the Otekura Formation is underlain by several kilometres of dominantly fine grained, deep water slope sediments, containing occasional sand and conglomerate filled channels similar to those here described in detail from the Otekura Formation. Such channels are inferred to form when a mass-transported sand, derived from failure higher on the slope, ploughs erosively into the sea floor. After their incision, the channels served for a short time as conduits for downslope transport of sediment, the redeposited deposits of which are found filling each channel. Both channel fills at Sandy Bay are capped by thin-bedded turbidites inferred to have overspilled from similar channels nearby on the slope.

INTRODUCTION

The role played by continental slope mass-failure in redistributing sediment to the deep sea was early stressed by Milne (1897). Later workers have emphasized the importance of transport via submarine canyons and channels, a comprehensive

0037–0746/79/0800–0475 $02.00

review being provided by Shepard & Dill (1966). Many ancient examples of submarine canyons and channels have now been recognized (see review by Whitaker, 1976), but detailed facies descriptions of their sedimentary fill remain rare. For the two slope channels described in this paper, it is argued that whereas the channels originated by a mechanism similar to that envisaged by Milne (1897), they then functioned as conduits that funnelled terrigenous sediment to deep water, as argued by, for example, Shepard & Dill (1966).

Geological setting

The Triassic and early Jurassic parts of the volcanogenic Hokonui Assemblage of the Rangitata Orogen are well exposed in the South Otago coastal sequence, southeastern South Island (Fig. 1). The Assemblage, at least 10 km thick at the coast, is

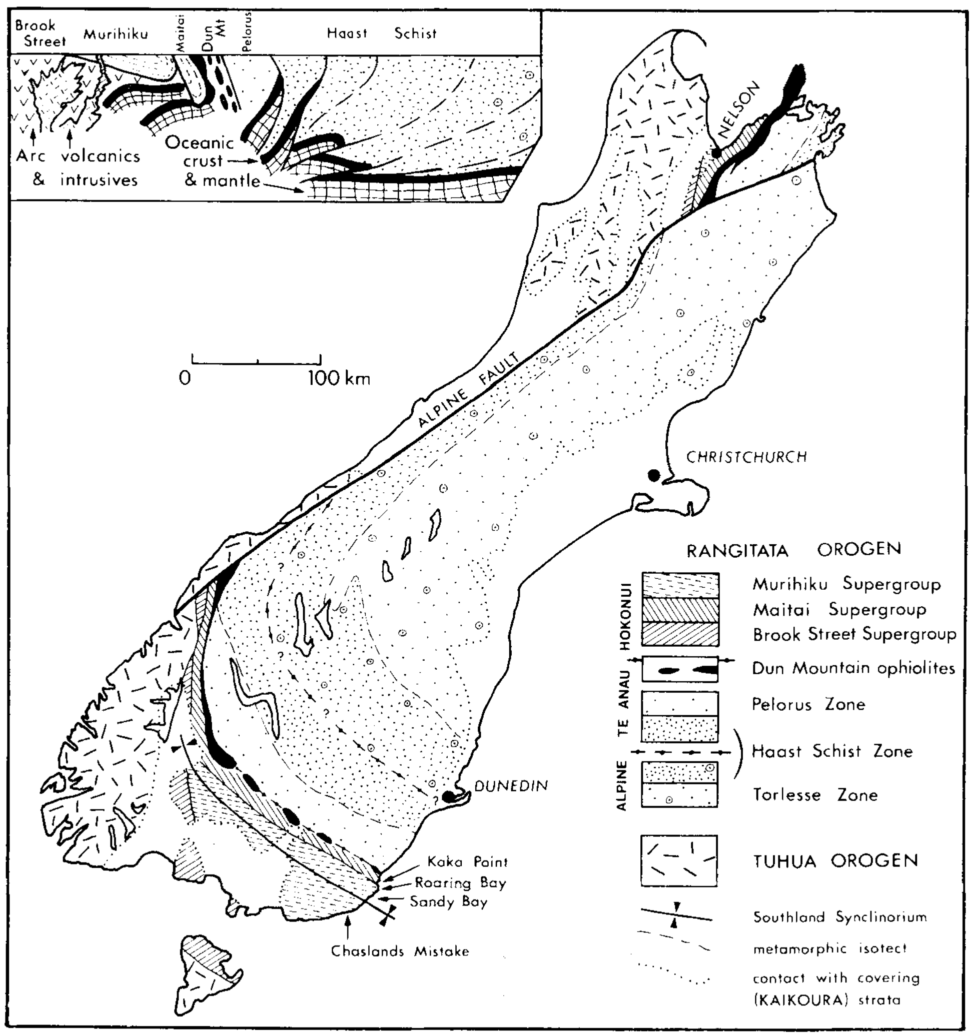

Fig. 1. Location map, showing regional geological setting of Sandy Bay. Inset, pictographic cross-section through the Rangitata Orogen, after Coombs *et al.* (1976).

gently folded into the regional Southland Synclinorium. The sediments on the north limb of the synclinorium are in sedimentary or fault contact with the inferred ancient ocean floor represented by the Dun Mountain Ophiolite Belt; they represent a regressive sedimentary suite of largely deep water origin, which finally shoaled to southerly derived shelf and non-marine facies in the early Jurassic (Carter *et al.*, 1978). The sediments on the south limb are in sedimentary or fault contact with the calc-alkaline volcanic and volcanogenic Brook Street Supergroup, which is assumed to represent the volcanic chain that flanked the Rangitata Geosyncline to the south and west (Grindley, 1958; Landis & Bishop, 1972). Though poorly known, south limb sediments that overlie the Brook Street Supergroup contain much more non-marine and shallow shelf sediment than their north limb stratigraphic equivalents (Fig. 2; cf. Speden, 1971).

The north limb coastal transect through the Hokonui Assemblage offers almost continuous cliff and wavecut platform exposure, particularly through the Murihiku Supergroup, and provides unparalleled opportunities to study deep marine sedimentation patterns within an arc-trench gap (or forearc basin, *sensu* Dickinson, 1974). This paper presents a description of the sedimentary facies associated with two submarine channels inferred to have been located on the upper trench slope of the Rangitata Geosyncline.

STRATIGRAPHY

Stratigraphical setting

The stratigraphic sequences provided by Park (1904), Mackie (1935), Campbell (1955), Speden & McKellar (1958) and Speden (1971) have been used as the basis for the sedimentological part of this study. Though only the uppermost 184 m of the Otekura Formation has yet been studied in any detail, the entire coastal sequence has been examined briefly as part of the background of this study. The stratigraphic setting of the Otekura Formation is summarized on Figs 2 and 3.

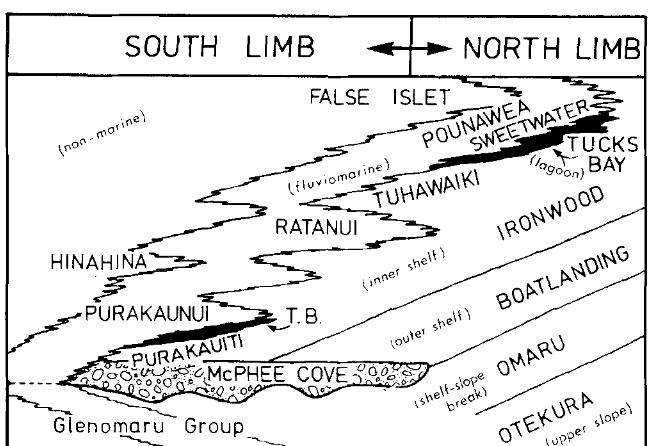

Fig. 2. Stratigraphical setting and nomenclature for the upper parts of the Murihiku Supergroup in the coastal exposures of the Southland Synclinorium. (Adapted from Speden, 1971.)

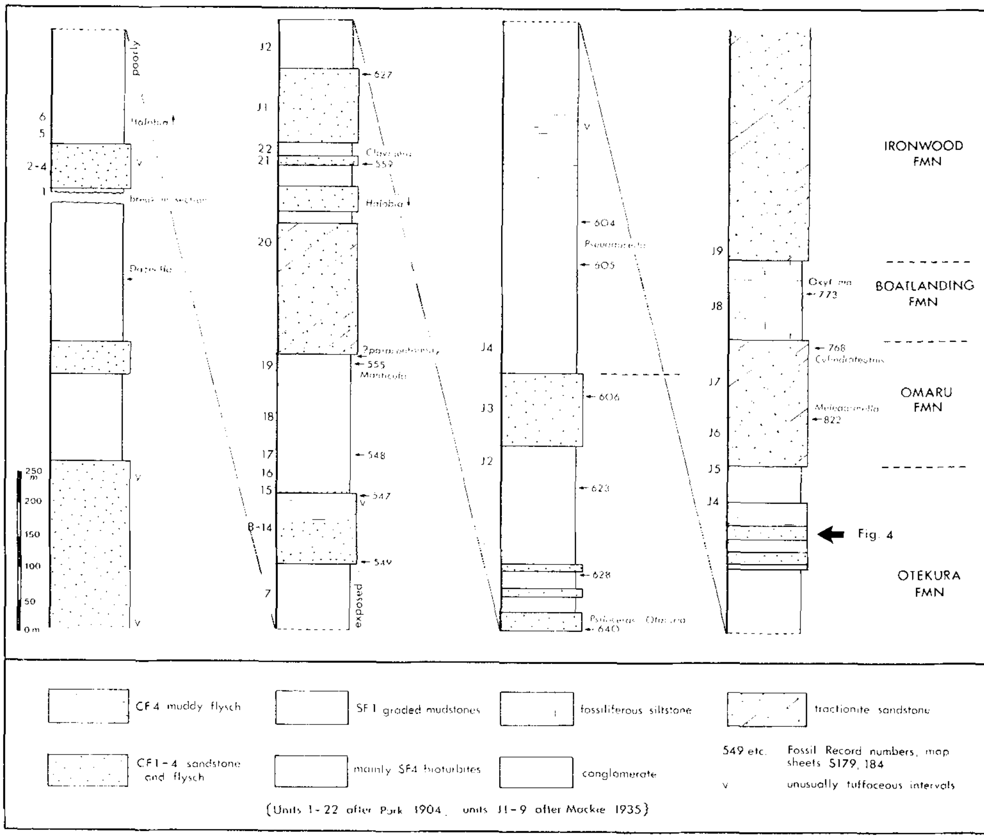

Fig. 3. Stratigraphical column of the lower and middle parts of Murihiku Supergroup exposed in the coastal section on the north limb of the Southland Synclinorium.

Otekura Formation

The Otekura Formation comprises some 600 m of fine-grained sediment, mudstone and siltstone predominating. With the exception of a zone of profuse *Pseudaucella marshalli* (Trechmann) in the middle of the formation, body fossils are generally absent. The presence of *Pseudaucella* indicates an early Jurassic ((?) Toarcian) age for the formation.

The Otekura Formation mudstones often have a flaggy or fissile appearance, due to the presence of spaced thin beds (generally less than 5 mm thick) of silt–very fine sand, and occasional 5–10 cm thick beds of tuff. Zeolitic and calcitic concretions are scattered throughout. Locally the mudstone–siltstone alternations disappear, the sediment appearing massive as a result of pervasive bioturbation; *Chondrites* is the dominant trace-fossil, and is abundant throughout the formation.

Most of the upper 184 m of the formation comprises a channelized flysch sequence, described in more detail below. The thin-bedded turbidites that occur at the top of the higher channel fill sequence (Fig. 4) become increasingly reworked by slumping and gentle bottom currents upwards, and pass gradationally into a few tens of metres of more massive mudstone with spaced, millimetre scale, wispy

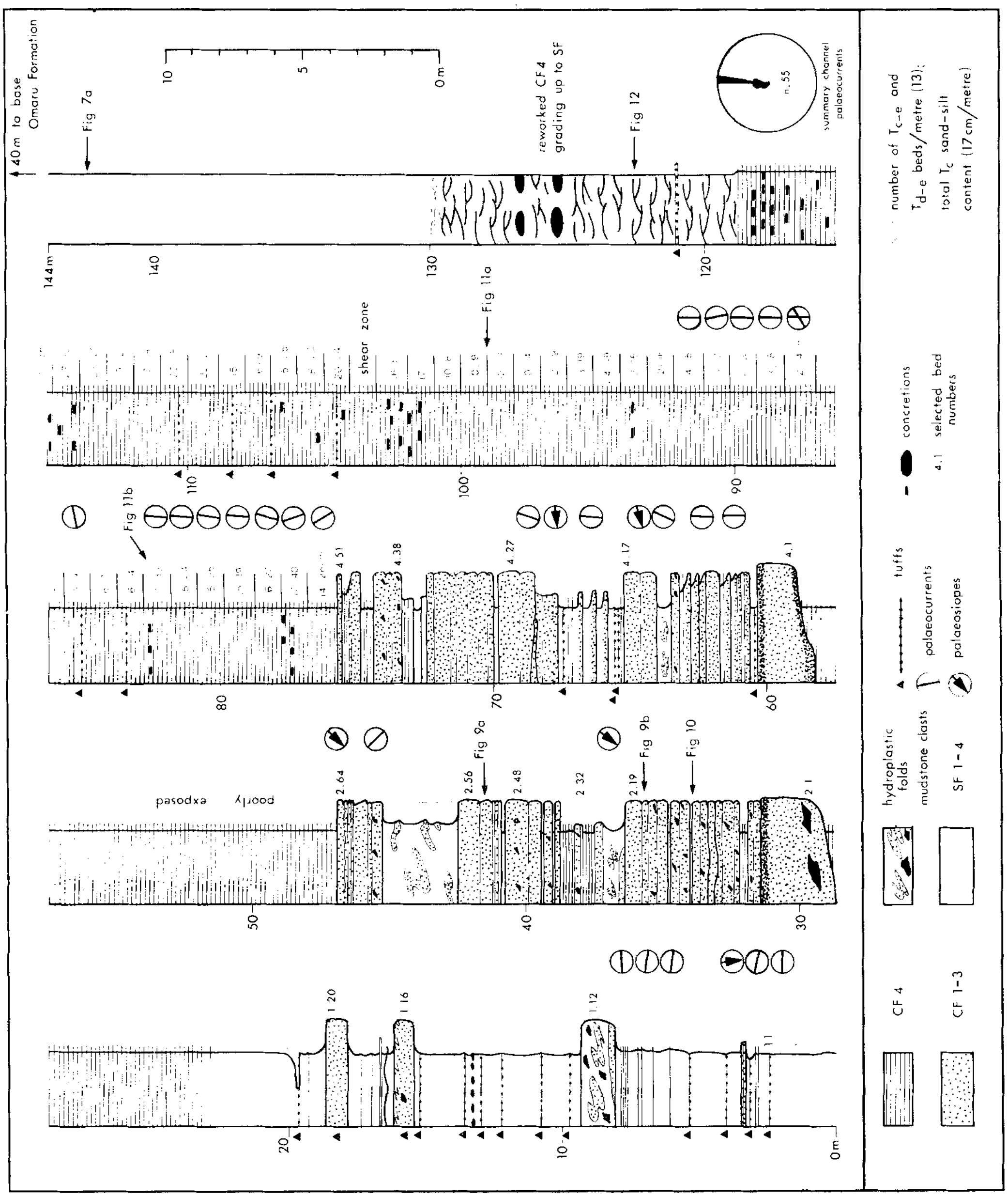

Fig. 4. Detailed section through the upper 184 m of the Otekura Formation, Sandy Bay, South Otago.

cross-laminae of light-coloured silt. Small-scale penecontemporaneous hydroplastic folds are common in these siltstones.

Underlying lithologies

Very similar lithologies to those described for the lower and middle parts of the Otekura Formation occur throughout the thick underlying sequence of Triassic and earliest Jurassic beds (Speden & McKellar, 1958). The sequence is predominantly fine-grained, with irregularly spaced occurrences of redeposited sandy sediment, occasionally conglomeratic, usually as channelized packets up to a few tens of metres thick. Trace fossils are abundant, particularly *Chondrites* and *Zoophycus*, and there are spasmodic occurrences of macrofossils (molluscs and brachiopods), some of which may be redeposited.

Overlying lithologies

Six formations have been mapped and described above the Otekura Formation by Speden (1971; see Fig. 2). The upper five formations represent a northwards advancing, regressive, shelf-nonmarine sedimentary wedge, culminating in the fluvial conglomerates of the False Islet Formation (Carter *et al.*, 1978). The sixth formation, the Omaru Formation, comprises 185 m of largely cross-bedded sandy and conglomeratic sediment. Zones of hydroplastically contorted sandstone also occur, together with intraformational conglomerates. Macrofossils, particularly molluscs, are conspicuously more common than in the Otekura and underlying beds; some are probably preserved *in situ*, whilst others are redeposited.

Stratigraphical interpretation

The sediments of the Murihiku Supergroup on the north limb of the Southland Synclinorium represent a regressive sedimentary wedge. Regional considerations suggest that the wedge was situated in a forearc setting, fed by a volcanic chain to the south and west, and connecting with a midslope basin and trench to the north and east (Coombs *et al.*, 1976; Carter *et al.*, 1978). Shallow water marine ('shelf') and non-marine facies are well developed in the Murihiku beds on the south limb of the synclinorium (Speden, 1971), and in the upper formations on the north limb (Fig. 2). The Omaru Formation has many of the characteristics to be expected in sediments that accumulated at a shelf-slope break (cf. Southard & Stanley, 1976). The Otekura Formation and underlying beds are interpreted as an upper slope facies for the following reasons: (1) they are immediately overlain by a regressive, shelf-type sequence, the base of which (Omaru Formation) is inferred to mark the shelf-slope transition (cf. Fig. 2); (2) they comprise a several kilometre thick sequence of mudstone, inferred to have been deposited below wave-base and including mud-turbidites; (3) the occasional coarser beds are generally redeposited, in contrast to the traction deposits of the overlying shelf sequence (cf. Fig. 3); (4) benthic fossils are sparse, again a contrast with the shelf sequence above; (5) regional interpretations, based on Waltherian sedimentary models, agree with points (1)–(4) above in suggesting a slope facies, but indicate more precisely that the environment was that of an upper trench slope (cf. Coombs *et al.*, 1976; Carter *et al.*, 1978).

The remainder of this paper describes two flysch-filled channels that occur towards the top of this inferred trench slope sequence.

SEDIMENTARY FACIES ANALYSIS

Slope facies association

The lower 23 m of the measured section (Fig. 4) comprise mudstone with minor traction-emplaced siltstone and redeposited siltstone or sandstone. Including for convenience some sediments that occur a little below the base of the section, four main lithofacies are distinguished.

Graded silty mudstones (SF 1) (Fig. 5)

This facies comprises green-grey and brown-grey beds from less than one to greater than 10 mm thick and with excellent lateral bed continuity and thickness. Beds have sharp bases, occasionally show micro-scours or induced micro-flames against the top of the underlying unit, and have well developed distribution grading. Each bed has the form of a couplet. The lower division is green-grey graded silt, with a maximum grain size of about 4–5 ϕ, and sometimes shows fine parallel lamination due to the alignment of mineral grains. The upper division follows, across a sharply gradational boundary, and comprises a darker, brown-grey mud (see SF 2 below). The relative thickness of the couplet divisions are variable, but commonly the mud

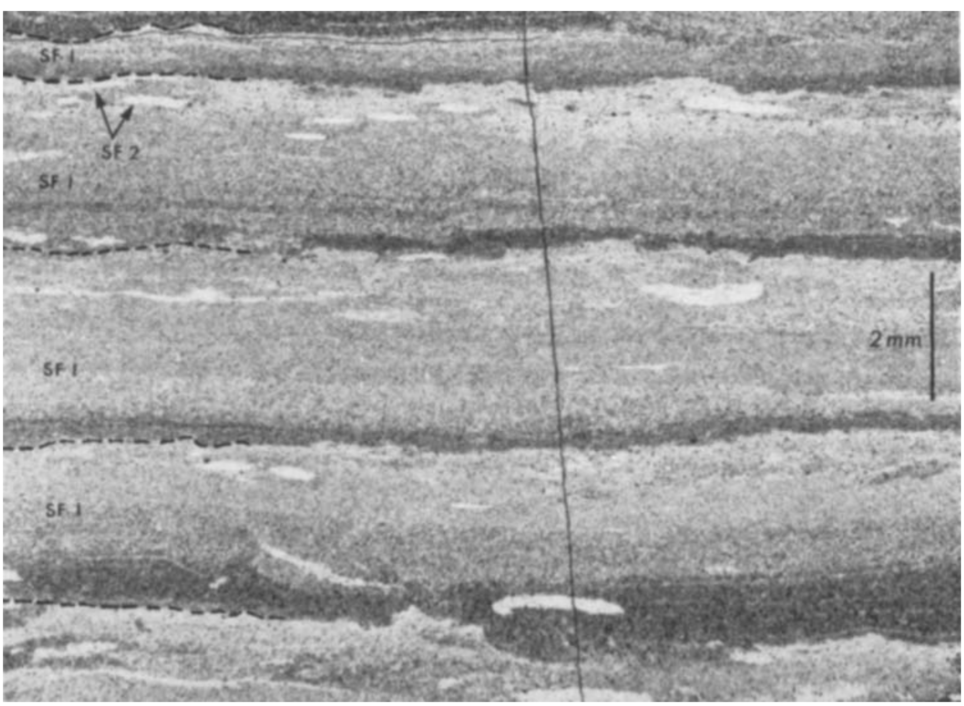

Fig. 5. Negative print from thin-section of SF 1 graded silty mudstones, 100 m below the base of unit 1 (Fig. 4), Sandy Bay. Thin layers of SF 2 mudstone occur at the tops of some beds, and within the *Chondrites* burrows.

has a thickness of 1–2 mm irrespective of the precise thickness of the associated graded silt division, which may range up to 10 mm or more.

Massive mudstones (SF 2) (Figs 5 and 11b)

This facies comprises beds or laminae of dark grey, massive mudstone. Individual units are generally one to several millimetres thick, rarely attaining thicknesses of a few centimetres. Occasionally, beds of this lithofacies contain elongated phosphatic concretions up to several tens of centimetres in longest diameter (cf. Fig. 11b). Some phosphatic layers contain well preserved calcispheres of biopelagic origin. The massive mudstone facies occurs as thin interbeds between other lithofacies, which it generally overlies with a sharp contact.

Rippled siltstones (SF 3) (Figs 6–7)

This facies comprises 2–10 mm thick beds of green or green-grey siltstone or, occasionally, very fine sandstone (Fig. 7). Individual beds have sharp bases, but the irregularly undulose (rippled) tops result in marked fluctuations in bed thickness from place to place. Individual beds can be seen to wedge out when traced several metres laterally. Many of the beds show well developed cross-lamination, which is often accentuated by foreset mud-laminae, or by individual foreset laminae being rich in disseminated opaque minerals (particularly common in this lithofacies).

Towards the top of the Otekura Formation, above the channel-fill sequences (cf. Fig. 4), SF 3 rippled siltstones occur intermixed with SF 1 graded mudstones (Fig. 6). Rather than forming discrete spaced beds, as in the typical occurrence, the rippled silts are here spasmodically intermixed with the background graded muds and silts as wispy and extremely discontinuous layers, some with internal cross-lamination,

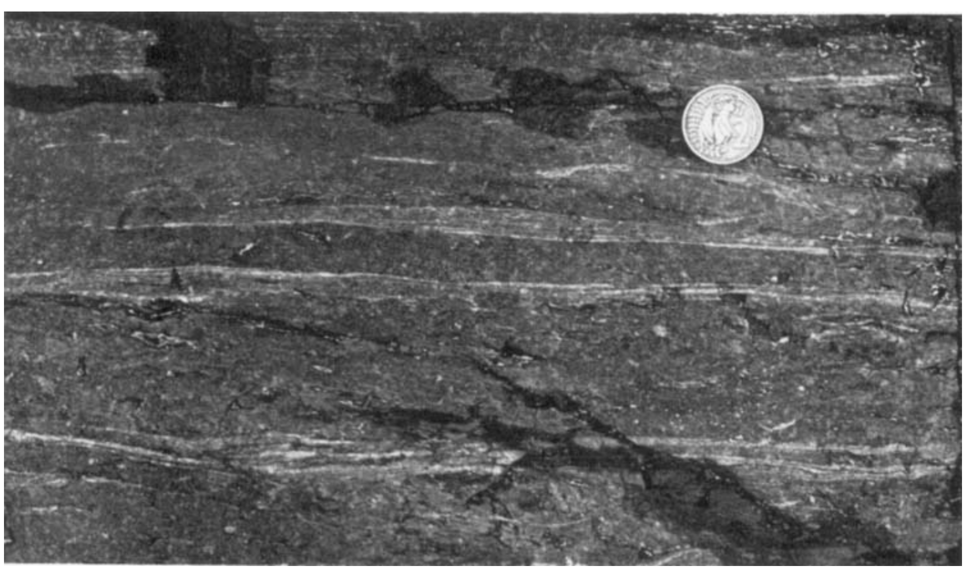

Fig. 6. Outcrop view of SF 3 rippled siltstones separated by SF 1–2 mudstones, 143 m above base of measured section (Fig. 4), Sandy Bay. Coin, 2·1 cm.

others as thin silt drapes over wavey bedforms on top of the underlying stratum. Lithofacies boundaries are difficult to locate precisely in this sequence, and future investigation may result in the recognition of a discrete lithofacies.

Bioturbated lithofacies (SF 4)

Mud-filled burrows of the *Chondrites* type are scattered throughout the siltstones and mudstones of lithofacies SF 1–3 (cf. Figs 5 and 7). They are most abundant towards the top of individual sedimentation units, i.e. directly below the inferred sea floor represented by the top of each sedimentation unit. Occasionally, *Chondrites* may be associated with larger deposit feeding traces, the combined organic activity having destroyed all fine sedimentation structures. It is convenient to recognize beds of this type, actually complex intermixtures of several SF lithofacies, as a discrete bioturbated lithofacies.

Interpretation of slope lithofacies 1–3

The SF 1 graded silty mudstones correspond to $T_{d\text{-}et}$, $T_{d\text{-}ep}$ and T_{et} sequences of the Bouma classification of turbidity current deposits (Bouma, 1962). Though they

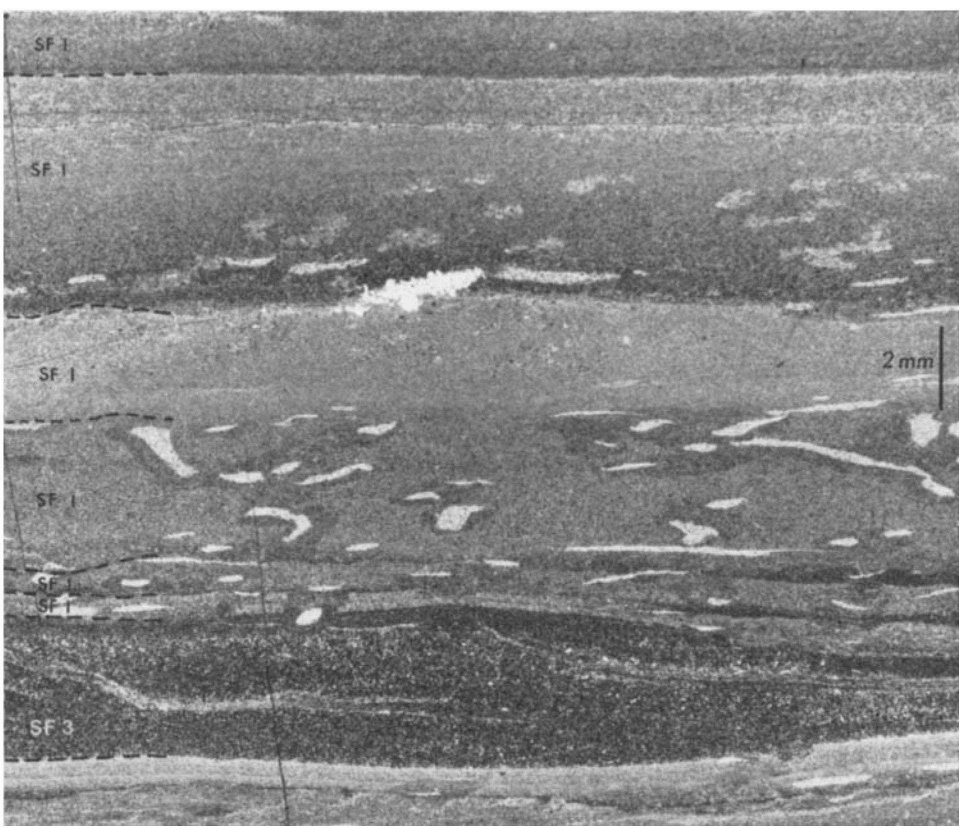

Fig. 7. Negative print from thin-section of SF 3 rippled siltstone, 100 m below the base of unit 1 (Fig. 4), Sandy Bay. (Note abundant *Chondrites* in overlying SF 1 graded mudstones).

might conventionally be regarded as 'distal' turbidites (e.g. Walker, 1967), lithofacies SF 1 muds in fact correspond closely to the expected deposits of mud-turbidites (nephelites) (Moore, 1970; cf. Le Pichon *et al.*, 1976), in which case they have no necessary 'proximal' equivalent. Additionally, lithofacies SF 1 may include truly distal turbidites of distant overbank or slope-basin plain origin. Similar facies have been interpreted as mud turbidites by Griggs, Carey & Kulm (1969), Piper (1972), Hall & Stanley (1973), and Hesse (1975).

Massive mudstones of lithofacies SF 2 were probably deposited directly from suspension, as inferred for similar modern muds (Griggs *et al.*, 1969). Where interstratified with mud-turbidites, these terrigenous or hemiterrigenous muds (Scholl & Marlow, 1974) would be classified as subdivision T_{ep} of the Bouma sequence (cf. Fig. 11b). The reason for the relative rarity of biopelagic remains in SF 2 mudstone is not clearly understood; though the main groups of calcareous plankton were not present in the early Jurassic, there were certainly contemporaneous siliceous plankters. The absence of radiolarian cherts is a general feature of the Hokonui Assemblage facies-belt (Landis, in Blake, Jones & Landis, 1974), perhaps suggesting that silica was resorbed during suspension settling, deposition or early diagenesis of the sediments. That calcareous plankton are preferentially preserved in early diagenetic (pre-compaction) phosphate concretions certainly suggests resorption of calcareous plankton into interstitial waters at an early stage of sediment burial.

Rippled siltstones of lithofacies SF 3 were emplaced from intermittent but persistent bottom currents. Initially the silt may have been introduced into the depositional area by turbidity currents, but it may then have been redistributed along slope by bottom currents. Sediments similar to those of lithofacies SF 3 have been described by Moore (1974; Eocene 'laminites' from DSDP hole 241 on the East African continental slope), and by Bein & Weiler (1976; amongst Cretaceous 'calcilaminites' from Israel), and tentatively ascribed to deposition by contour currents (cf. Hollister & Heezen, 1972). It is therefore considered possible that SF 3 siltstones were deposited from contour currents. As, however, Hesse has noted (1975, p. 411), it may be very difficult in particular cases to 'distinguish between a (mud) turbidite and a "contourite", because both may be expected to show the effects of hydraulic sorting and the depositional mechanism in both cases may be very similar'.

Channel facies association

The upper parts of the Otekura Formation contain two conspicuous packets of thicker bedded sandstones, or flysch (Figs 4 and 8), which are inferred to have been deposited in small-scale slope channels (cf. Carter *et al.*, 1978). The evidence for the presence of channels, and descriptions of the four main flysch lithofacies they contain, are given below.

Evidence for the presence of channels

The generally poor inland outcrop precludes direct observation of channel walls. Nonetheless, the following facts suggest that flysch sequences such as those of Fig. 4 were deposited in channels, the lateral extent of which ranged up to several hundred metres and which had depths between a few and several tens of metres.

(1) Aerial photographs, and detailed mapping (Boles, 1974), show that the sandy

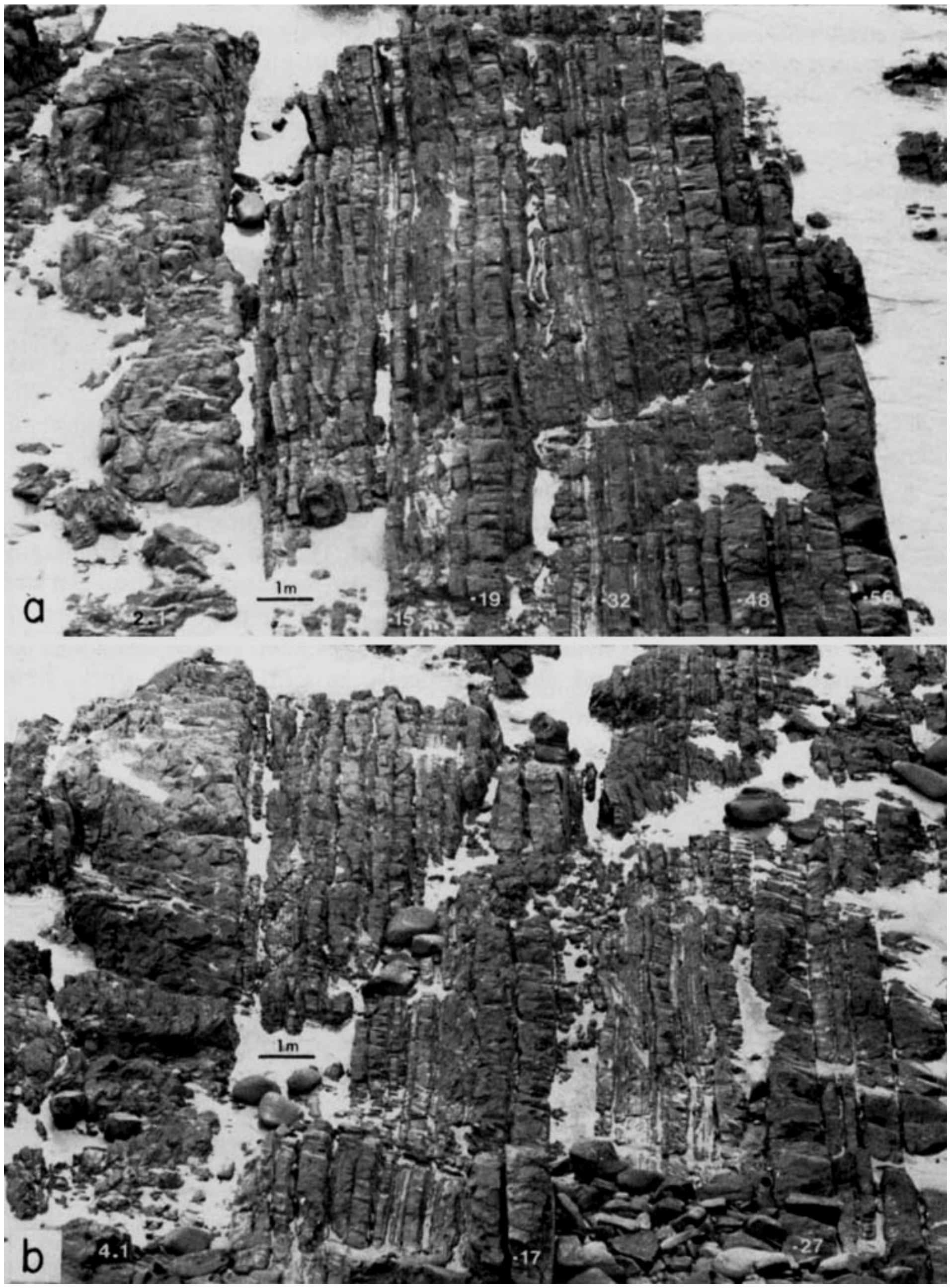

Fig. 8. Outcrop views of two flysch-filled channels, (a) unit 2; and (b) unit 4, upper Otekura Formation, Sandy Bay (cf. Fig. 4). Selected beds numbered; youngs to right.

flysch sequences occur as lensoid bodies within an SF 1–4 argillite background. The lenses range from one hundred to several hundred metres long and are generally a few tens of metres thick.

(2) In outcrop, as at Sandy Bay, the sandstone at the base of each packet of sandy flysch has an erosive base with relief of up to a few metres; basal sandstones also commonly contain large rip-up clasts.

(3) Both packets of sandy flysch at Sandy Bay contain redeposited sediments of various types (CF 1–4 below), for which palaeocurrent indicators show that transport took place to the north or northeast, i.e. down the regional palaeoslope (cf. Figs 1 and 4). Since the flysch packets are not sheet-like (points (1) and (2) above), they were presumably emplaced in channels.

In view of the current popularity of submarine-fan models for the interpretation of flysch sequences, it should be stressed that the flysch sequences of Fig. 4 occur in a thick sedimentary background that is dominated by argillite, and that typical fan sequences or rhythms are absent (cf. Fig. 3).

Massive sandstones (CF 1) (Fig. 8)

A thick bed of massive sandstone occurs at the base of each flysch sequence (e.g. bed 4.1, Fig. 8b). The base is sharp, erosive into the underlying muddy sediments, and undulating over lateral distances of several metres. The top of the bed is laterally more nearly planar, but with minor undulations of decimetre wavelengths. The bed at the base of the lower channel (bed 2.1) contains large angular clasts of rip-up mudstone and siltstone, some hydroplastically contorted. CF 1 sandstones may be slightly distribution graded, and characteristically have a 20–40 cm thick, light

Fig. 9. Thick bedded CF 2 sandstones (Beds 2.18–2.19) in the middle and upper parts of the lower channel fill, unit 2, Sandy Bay (cf. Fig. 4); hammer, 85 cm.

weathering, slightly finer grained selvedge at both top and bottom of the bed. The two CF 1 sandstones within the measured section have thicknesses varying between 150 and 350 cm, and aerial photographs suggest a lateral extent of 1–2 km. Similar lithofacies occur throughout the Murihiku Supergroup sediments that underlie the Otekura Formation, where CF 1 sandstones up to 12 m thick occur.

The lithofacies is similar to Facies A of Mutti & Ricci Lucchi (1972), and to the Massive Bedded Sandstone Facies of Collinson (1970).

Thick-bedded sandstones (CF 2) (Figs 8 and 9)

The main part of each channel fill comprises somewhat irregular but laterally extensive beds of massive sandstone between 5 and 80 cm thick, with only minor intervening mudstones (Figs 8 and 9). Sandstone bed bases are sharp. Grading is not conspicuous except over the top few centimetres of each bed, where abundant lenses and laminae of silt–mud may occur. These are generally streaked-out or folded by hydroplastic shear, the transition zone often being marked by irregular, chippy jointing (Fig. 9, arrowed). Many beds have an undulatory top, of amplitude up to 10 cm and wavelength up to 50 cm, over which the next bed is then apparently draped. Faint parallel or wavey partings occur in the upper half of some beds; others contain well developed dish-structures throughout, reflecting fluidization on deposition.

The lithofacies is similar to Facies B of Mutti & Ricci Lucchi (1972).

Graded sandstones–siltstones (CF 3) (Fig. 10)

Normally graded, 10–35 cm thick, tabular sandstones–siltstones of turbidite type also occur in the main channel fill, but are uncommon. Bouma features other than grading are rare, though sometimes a zone of wavey ripple cross-bedding occurs around the junction between the sand and silt parts of each bed ($T_{c\text{-}d}$ level). Grain-size decreases rapidly but gradationally across this zone, diagenetic recrystallization often causing a spuriously sharp cementation break in each couplet near the sand–silt contact (Fig. 10). Sandstone bases are sharp, sand:silt ratios are of the order of 3+:1, and some beds have zones of angular mudstone clasts up to 25 cm in diameter, usually towards the top of the sand subdivision.

The lithofacies corresponds to the immature Facies C of Mutti & Ricci Lucchi (1972), and Ricci Lucchi (1975).

Thin-bedded graded sandstones–mudstones (CF 4) (Fig. 11)

Each channel sequence is underlain and capped by a zone of extremely regularly thin-bedded sandstone/siltstone and mudstone of 'distal' turbidite aspect. Sandstones have very fine to fine sand grain size, are 1–3 cm thick (rarely up to 6 cm), and pass up gradationally into siltstone and mudstone of the upper part of each couplet; sand/silt:mud ratios are of the order of 1:2–4. Sandstones bases are generally planar with abundant sole markings, notably (1) shallow groove and drag marks; (2) occasional oval sole marks up to 5 cm long and with coarser grained fill (up to medium sand) than the grain size of the overlying sand; and (3) infillings of irregular biogenic depressions on top of the underlying mudstone or siltstone. Bouma sequence features are common, with T_c low angle, climbing, ripple cross-lamination conspicuous in nearly all beds; stoss side laminae are usually eroded, and there are usually one to

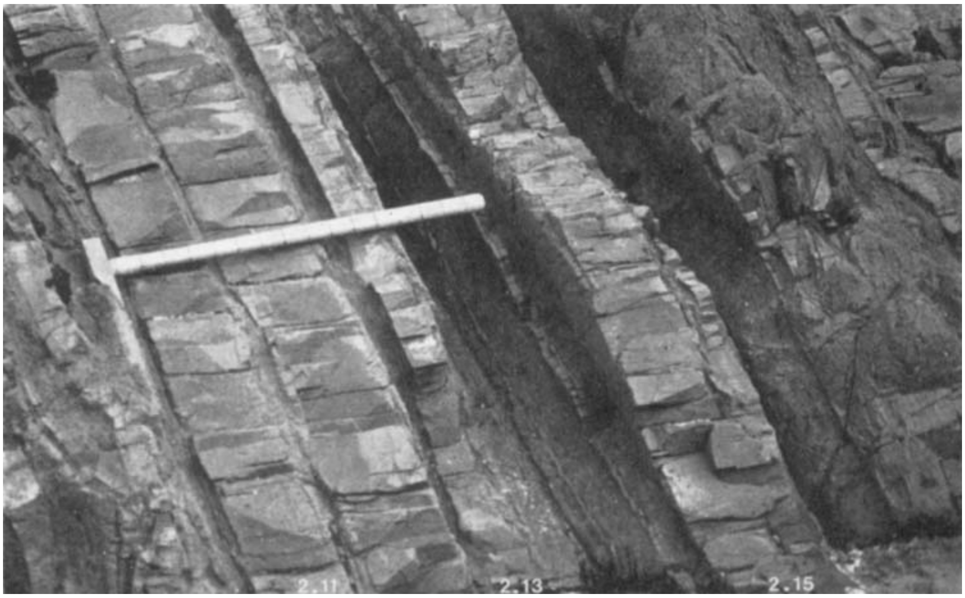

Fig. 10. Tabular graded sandstones-siltstones of CF 3 in the lower part of the lower channel fill, unit 2, Sandy Bay (cf. Fig. 4); hammer, 85 cm. Section youngs to right. Note that both beds 2.10 and 2.16 are hydroplastically deformed.

two (maximum three) sets of cross-beds in each coset. T_b parallel laminations are present in some thicker beds. T_d laminations are common, and in some cases are thrown into gentle syndepositional hydroplastic folds. Most beds correspond to Bouma T_{c-e} sequences, with T_{b-e}, $T_{a,c-e}$ and T_{d-e} sequences present in lesser numbers. Bouma T_{et} and T_{ep} are distinguishable by subtle colour and grain-size differences in some beds; occasional T_{ep} layers (lithofacies SF 2) contain small, ovoid, pre-compaction phosphate concretions.

A study of the 43 m thick sequence of lithofacies CF 4 which overlies the upper channel fill suggests a thinning-fining upward trend. Particularly, (1) an increase in the number of sandstones per metre from twelve to seventeen per m at the bottom to seventeen to thirty per m at the top (Fig. 4); (2) a concomitant decrease in thickness of the sandstone layers, from 2 to 3 cm thick at the bottom to 0–1 cm thick at the top; (3) a concomitant increase in the number of T_{d-e} turbidites higher in the section.

Lithofacies CF 4 is equivalent to Facies D2 and D3 of Mutti & Ricci Lucchi (1972) and Ricci Lucchi (1975).

Interpretation of slope channel lithofacies 1–4

Slope channel lithofacies 1–4 are entirely redeposited sediments, with the exception of minor amounts of SF 2 suspension emplaced mudstone that occur between sand beds, particularly in lithofacies CF 4. Many writers have described and interpreted similar lithofacies before, particularly from inferred ancient submarine fan or flysch basin sequences (see especially Ricci Lucchi, 1975, and the many papers and references in Dott & Shaver, 1974).

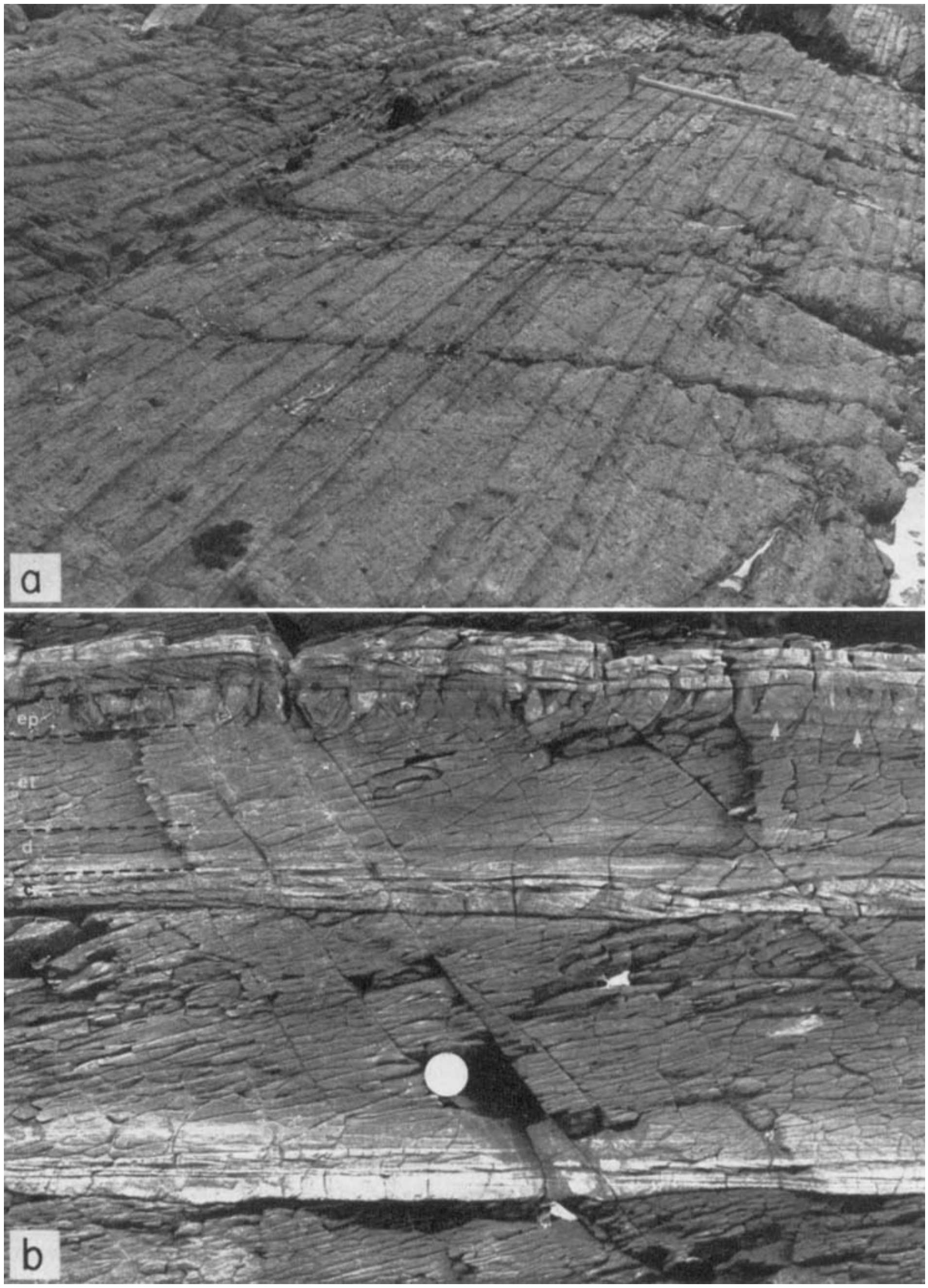

Fig. 11. Thin-bedded graded sandstones-mudstones of CF 4, above the upper channel-fill sequence (cf. Fig. 4), Sandy Bay. (a) 23 metres above the base of unit 5, hammer 85 cm. Beds young left. (b) Detail of two Bouma T_{c-e} sequences, one with phosphatic concretions growing within e_p (arrowed); coin 2·1 cm.

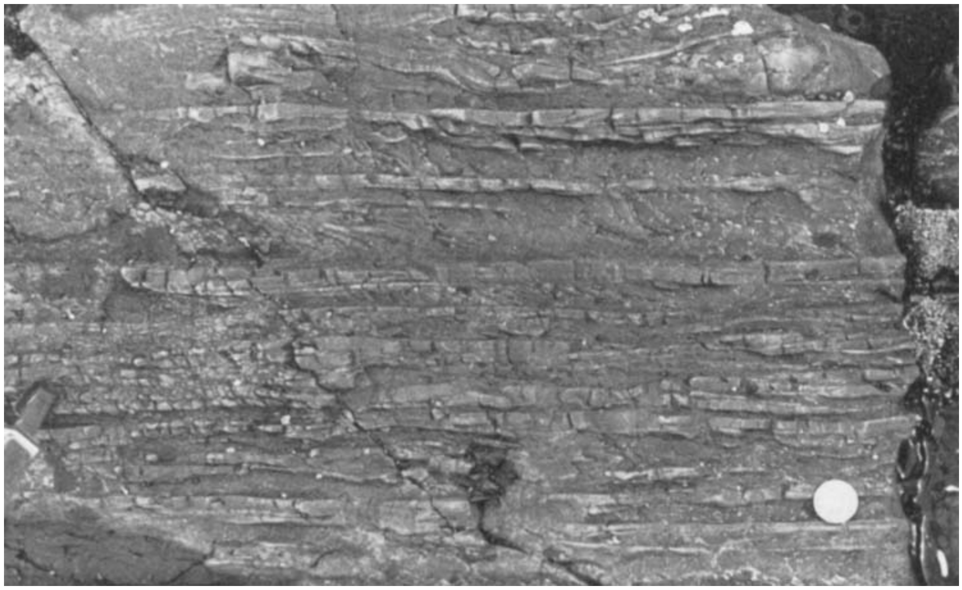

Fig. 12. Irregularly slumped (and ? traction reworked) thin-bedded CF 4 sands and muds, 123 m above the base of measured section (Fig. 4), Sandy Bay; coin 2·1 cm. Contrast with Figs 7 and 11.

The most obvious interpretation for lithofacies CF 1–4 is that they were deposited from turbidity currents. This interpretation is favoured for facies CF 3 and CF 4, both of which are similar to turbidites as described and figured by many previous authors. The poor development of Bouma sub-divisions in CF 3 graded sandstones is probably a reflection of the well sorted nature of the flow source material; the spasmodic and rare occurrence of CF 3 beds amongst the other channel fill facies suggests that true turbidity currents were not the typical mode of transport through the channels here described (see below). Facies CF 4, on the other hand, displays abundant Bouma features, and these thin-bedded turbidites are closely similar to inferred overbank turbidites of earlier authors (e.g. Nelson & Nilsen, 1974; Carter & Lindqvist, 1977). An alternative explanation might be that facies CF 4 was deposited within the channel, but from turbidity currents of lessening size during a phase of channel abandonment and filling.

Interpretation of facies CF 2 thick-bedded sandstones is rendered difficult by the absence of detailed sedimentary structures and by the not uncommon bed disruption caused by dish or other liquefaction structures. The beds are not typical turbidites, however, because (1) they do not contain Bouma features; (2) they are either ungraded or extremely poorly reverse or normal graded; (3) they are rarely tabular and may be laterally discontinuous over distances of several tens of metres; and (4) they were probably at least sometimes deposited with an undulating top surface, over which succeeding beds are draped. Similar redeposited sediments have been interpreted by earlier authors either as 'proximal turbidites', or as 'grain flows'. For reasons discussed more fully elsewhere (Carter, 1975), the interpretation preferred here is that these beds were deposited from an inertia-flow carpet at the base of a fluxoturbidity current, i.e. they are fluxoturbidites in the original and strict sense.

Similar arguments apply also to the transport and deposition of massive sandstones CF 1. The presence of large rip-up clasts and the incised basal contact to CF 1 beds establishes their high-energy and erosive emplacement. Further, for the well exposed examples in the New Zealand coastal section, and for others described in the literature (e.g. Lowe, 1972a), the stratigraphical position of CF 1 sandstones is either as isolated beds in a mudstone background, or at the base of a sandstone packet or 'rhythm'. This fact is unlikely to be coincidental and, at least for the New Zealand examples, appears most consistent with the emplacement of a large mass of gravity driven sediment onto a soft seafloor, often but not always with the result that a downslope channel became established. The precise mechanism of transport for CF 1 sandstones is necessarily a speculative inference; the conservative position taken here is that they were deposited from inertia flows (cf. Sanders, 1965; Carter, 1975).

PALAEOSLOPES

Current indicators

Flow-indicators are not common in the thicker bedded sandstones of lithofacies CF 1–3 within the channel-fill sequences. Those that occur, largely drag marks on the base of CF 2 sands, indicate north–south alignment of the depositing current (Fig. 4). Similarly aligned sole marks are common on the bed-bases of lithofacies CF 4 overbank turbidites. The associated cross-bedding in all cases identifies the current-direction as up-dip, i.e. implies emplacement from northerly flowing currents (Fig. 4, summary rose diagram). Such a transport direction is in accord with the regional sedimentary evidence (Speden, 1971; Carter *et al.*, 1978).

Occasional turbidites of lithofacies CF 3 or CF 4 occur in the 30 m of sediment exposed below the lower Otekura Formation channel (Fig. 4). Several beds have well developed sole marks which, together with associated ripple cross-beds, indicate westerly directed transport, i.e. turbidity currents flowing along slope at right angles to the main downslope channel systems. The presence of such longslope turbidites suggests as one possible interpretation the presence of a small upper trench-slope basin along the axis of which the turbidity currents were flowing. (Current directions are not easy to measure in contourite lithofacies SF3, but are apparently broadly consistent with an east–west slope alignment).

Slope indicators

Indications of local submarine slopes occur throughout the Murihiku Supergroup. Most common are hydroplastically deformed horizons up to a few tens of centimetres thick, with folds outlined by contained silt or sand layers. Packets of parallel bedded mudstone and siltstone several metres to several tens of metres thick, in which strike is subtly but definitely distinct from that of the sediments above and below, are visible on aerial photographs of Murihiku Supergroup sediments below the Otekura Formation; these larger scale features probably represent substantial submarine slump packets (cf. Piper, Normark & Ingle, 1976).

Within the Otekura Formation small-scale hydroplastic deformation is commonest towards the top of the upper zone of CF 4 overbank turbidites (Fig. 12) and in the overlying muds and silts of SF 1–3 (Fig. 4), consistent with a slightly steeper sea-floor slope just below the shelf-slope break (cf. Fig. 2).

Slightly larger scale hydroplastically contorted horizons are not uncommon within the channelized thicker bedded sands of CF 2–3 (Figs 8a and 10). The affected zones are generally 40–60 cm thick, and sometimes pass laterally into undeformed beds. Within the deformed zones individual sandstone bands are 5–15 cm thick and may (1) wedge out sharply laterally into the surrounding mudstone; (2) bifurcate laterally into sharply bounded 1–2 cm thick sandstone layers (? sandstone 'sills') separated by mudstone; (3) be hydroplastically deformed into recumbent folds; and (4) be increasingly disrupted into angular blocks of sandstone set in a siltstone matrix, ultimately passing into a fully mobilized flowite or fluxo turbidite type bed (cf. Schlager & Schlager, 1973). The normals to the fold axes in hydroplastically deformed horizons are variably orientated, though usually with a component down regional slope (Fig. 4), suggesting the folds may represent a response to channel flank slopes.

A SLOPE-CHANNEL SEDIMENTARY MODEL

Channels of similar size and containing similar types of redeposited sedimentary fill to those of the Otekura Formation are now quite widely known. Most described occurrences, however, are either from inferred active submarine fan complexes (e.g. Piper & Normark, 1971; Carter & Lindqvist, 1975) or else represent major shelf-incising structures of considerable size and maturity (see many references in Whitaker, 1976). The evidence is clear that the Otekura channels themselves were not part of an active fan, even though they may have fed sediment to such constructional features further down their palaeoslope; rather, the channels are situated amidst a considerable thickness of submarine slope type sediments. Slope channels and their fills have previously been inferred in the Carboniferous of England (Walker, 1966; Collinson, 1970), the Devonian of U.S.A. (Hall & Stanley, 1973), the early Cainozoic of France and Poland (Stanley & Unrug, 1972), and the early Pleistocene of New Zealand (Lewis, 1976). The Otekura channels show more clearly than these earlier described examples a repetitive pattern of channel cutting and filling from which can be proposed a facies model which may have a more general applicability (cf. the model for fan channel sedimentation of Normark & Piper, 1969).

The formational stage

The CF 1 massive sandstones that occur at the base of each channel fill are inferred to have been emplaced by inertia-flow. They differ from other similar intra-channel facies, such as those of Lewis (1976) and Lowe (1972a, Fig. 4), only in being better sorted and in not containing exogenous conglomerate fractions, differences that are probably entirely due to local source conditions. The nature of the bottom and lateral contacts of CF 1 sandstones (cf. Fig. 3 in Collinson, 1970), and the fact that they may contain large rip-up clasts, establishes their erosive nature. It seems likely, therefore, that these slope channels became established when particularly erosive inertia-flows ploughed deeply into the substrate at some point after mobilization (cf. Lewis, 1976). That slope sediments are soft enough for this to be a feasible method of channel formation has recently been confirmed in an elegant experiment by the crew of the research submersible *Archimèdè* (Le Pichon *et al.*, 1976). A similar argument for the origin of slope channels and of at least some submarine canyons has been advanced by Hall & Stanley (1973) and Carter & Lindqvist (1975).

Lowe (1972b), in proposing a slope channel facies model based on examples from the Late Cretaceous of the Sacramento Valley, argued that the formative stage 'is not represented by channel deposits' and that 'the fill documents only the waning stages of channel evolution'. An alternative view, preferred here, is that the cutting event may leave its sedimentary deposit as a lining to the bottom of the channel; often in the form of a thick flowite of some type, as inferred for CF 1 sands within the Otekura Formation channels, and for many similar sands or conglomerates described in the literature.

Not all flowites of CF 1 type necessarily erode deeply enough to establish active channels. Examples of CF 1 massive sandstones seen in the Murihiku Supergroup below the Otekura Formation (e.g. beds 1.16, 1.20 of Fig. 4), as well as many similar 'grain flows' or 'debris-flows' reported in the literature, have erosive bases and yet are overlain directly by the same fine grained sediment types that they incise, rather than being followed by a typical active channel fill sequence (see below).

The functional and filling stages

Where CF 1 sandstones are followed by CF 2 fluxoturbidites and/or CF 3 turbidites, as in the two occurrences in the upper Otekura Formation (cf. Fig. 4) and at least one of the cases described by Collinson (1970) (Fig. 3), a slope channel must have existed down which these varied flows were passing and in which they deposited their more proximal phases. Discounting the basal CF 1 bed, the CF 2–4 fill of the channels reveals no particular thickening or thinning upward trend; rather, bed thicknesses within the channel fill apparently vary spasmodically, and the only obvious intra-channel organization is the presence of 2–5 m thick alternating packets of sandy CF 2–3 beds and thinner, muddier CF 4 beds (Figs 4 and 8). These spasmodic intra-channel variations between thin-bedded muddy flysch and thick-bedded sandy flysch might be interpreted (1) as reflecting the lateral migration of thalweg and thalweg overspill areas within the main channel (cf. Chough & Hesse, 1976), or (2) as due to irregularly fluctuating flow sizes, or strengths, over the life of the channel.

Similarity to fan-channel fills

There are obvious similarities between the channel-fill sequences described here from inferred slope channels and the patterns considered typical of submarine fan channels (e.g. Normark & Piper, 1969; Ricci Lucchi, 1975). In detail, however, there are significant differences, the most important of which is the absence of any clear thinning-fining upward cycles in the slope channel fills. Though a gross thinning upward trend is displayed by the sequence CF 1 (base of each channel), to CF 2–3 (middle) to CF 4 (top), in detail the main part of the channel fill sequence displays irregular inter-bedding of the various facies (cf. Fig. 8). A second difference is the presence of only a relatively small number of turbidites *sensu stricto* in the main fill of the channel. That most of the beds are rather of flowite-fluxoturbidite type is consistent with the inferred palaeogeographic position of the Sandy Bay channels, on an upper slope not far below the shelf edge. Presuming the flows to be derived by gravity failure at and around the shelf edge, it could in fact have been anticipated that the fill of such upper slope channels would consist largely of deposits from relatively immature forms of mass-transport, since most flows would still be taking off or accelerating in such a setting.

Abandonment

The CF 2–3 sandy flysch fills of the Otekura channels are followed abruptly by some thickness of extremely regularly bedded T_{c-e} turbidites of lithofacies CF 4 (Fig. 4). These thin-bedded turbidites are closely similar to the levee or overbank turbidites recently described by Mutti & Ricci Lucchi (1972), Ricci Lucchi (1975), Nelson & Nilsen (1974) and Carter & Lindqvist (1977). Their presence suggests that after each channel was filled, the locus of downslope transport shifted to a laterally adjacent channel; overbank turbidites from such a laterally adjacent channel may then cap the older channel fill sequence (cf. Normark & Piper, 1969; Carter & Lindqvist, 1975).

CONCLUSIONS AND SUMMARY

The Otekura Formation at Sandy Bay represents upper trench-slope environments of deposition. The background sediment type comprises mud turbidites interspersed with terrigenous or hemiterrigenous mudstones deposited from suspension, and with thin layers of siltstone possibly deposited from geostrophic currents.

The slope sequence contains occasional flysch-filled channels. These channels have an irregular fill of redeposited sediment, including flowites, fluxoturbidites and turbidites, and are associated with appreciable thicknesses of thin-bedded overbank T_{c-e} turbidites. The slope into which these channels were cut had an east–west strike, with sediment transport taking place downslope in a northerly direction.

A model is proposed for the development and filling of slope channels. Channels are formed when the seafloor is cut by a substantial inertia-flow or fluxoturbidity current, the massive sand deposit of which may be preserved as a lining to the channel. The channel then serves as a locus for downslope sediment transport by fluxoturbidity and turbidity currents, deposits from which form the main part of the channel fill. The irregular interbedding of thicker bedded fluxoturbidites, turbidites and thin-bedded 'overbank' turbidites, all within the main channel fill, is consistent with the lateral migration of a main thalweg within the channel. Channel sequences are sometimes capped by thick deposits of overbank T_{c-e} turbidites, derived by overspill from adjacent channels, and then by mudstones and siltstones of the general slope surface distant from an active channel.

ACKNOWLEDGMENTS

Grateful acknowledgment is made to the Otago University Research Grants Committee for financial support for fieldwork. Mr John Pillidge is thanked for his skilled preparation of thin-sections, particularly of the finer-grained lithologies; Dr A. R. Reay for X-ray diffraction determination of the phosphate nodules; and Drs R. J. Norris, E. Mutti, R. G. Walker and J. D. Collinson for their critical readings of the manuscript. I thank also Associate Professor J. D. Campbell and Dr E. R. Force for permission to include the occurrence of *Daonella* in Fig. 3.

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(*Manuscript received* 7 *March* 1978; *revision received* 21 *December* 1978)