

# Glacial sediment provenance, dispersal and deposition, Vestfold Hills, East Antarctica

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**Abstract:** Field observations involving landform evaluation and the physical characterization of sediments, combined with a detailed analysis of the spatial distribution of bedrock and sediment geochemical patterns, suggests a limit to the glacial transport half-distance of *c.* 3 km in Vestfold Hills. Four morphologically distinct glacial deposits were sampled (small debris ridges, large debris ridges, debris drapes and valley fills) on the basis of field geometry. These landforms were subsequently distinguished by grain size, mineralogy and geochemistry. Since there are no nunataks south of Vestfold Hills, all debris is derived subglacially and sedimentological differences are attributed to the physical weathering of preglacial surfaces in Vestfold Hills and fluvial winnowing during deposition. Given that thrust geometries may occur in large debris ridges, glacial transport distances were short, and fluvial sorting of sediments was an important mechanism, reconstructions of glacial histories based on the stratigraphy of deposits in Vestfold Hills should be made with caution.

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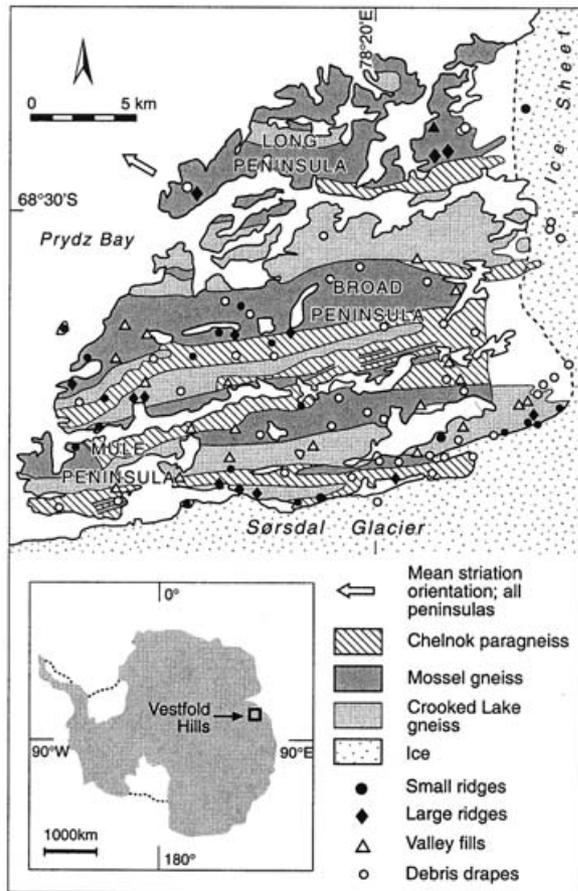
## Introduction

Surface sediments of Vestfold Hills, East Antarctica (Fig. 1), preserve a rich repository of information regarding their provenance, transport mechanisms and depositional processes. Several relatively abundant types of erratics that are not found in the bedrock of Vestfold Hills, including a distinctive banded iron-stone, indicate that at least some sediments have a provenance further to the east beneath the ice sheet. The relative contribution of distant versus local sources to sediment volume is unknown, and yet those sources control the physical and chemical characteristics of the sediment.

One way of determining whether sediment is derived from distant or local sources is to examine glacial dispersal patterns. As ice crosses a lithological boundary, the concentration of the former substrate as debris entrained in the ice, and deposited subglacially as till, decreases at an exponential rate (Krumbein 1937, Shilts 1976, Puranen 1990). An index of that decline is the glacial 'transport half-distance' - the distance over which half of the debris is lodged as basal till and replaced with newly eroded sediment or bedrock (Gillberg 1965). As an example, the transport half-distance for the Fennoscandian Ice Sheet lies in the range 1–10 km (Puranen 1990 and references therein). Glacial transport half-distance is a statistical estimate of the transport and deposition of individual particles, and is described by Puranen (1990) as the product of till depth and the sum of basal sliding rate and half of the ice creep rate, divided by the rate of substrate lowering.

None of these parameters have been measured for the ice sheet at Vestfold Hills, so an *a priori* calculation of glacial transport half-distance is impossible. However, it is possible to provide a semi-quantitative estimate of transport half-distance and thus sediment provenance, via an empirical investigation of till geochemistry. The pattern of glacial dispersal down-ice of 'point source' chemical anomalies variously forms ribbons, fans or plumes (Dyke & Morris 1988, Puranen 1990) depending on the geometry and orientation of the substrate and anomaly, substrate erodibility, rate and direction of basal sliding and internal creep of the ice. At the regional scale, plumes may also form down-ice of linear boundaries in areas of ice streaming (Boothia-type dispersal trains, *sensu* Dyke & Morris 1988). In contrast, at the kilometre scale, a sheet-like dispersal pattern may be expected down-ice of kilometre long linear boundaries that strike normal to ice flow direction.

The aims of this study were twofold. Firstly, to determine whether sediments of Vestfold Hills are derived from local or distant sources, through analyses of the chemical characteristics of glacial sediments across lithological boundaries. Secondly, to determine whether or not the sediments in the glacial depositional landforms could be distinguished on the basis of their physical and chemical attributes. In this way, aspects of the sediment dispersal pathways and depositional environments could be inferred, and the genesis of the deposits determined.



**Fig. 1.** Vestfold Hills showing sediment sampling sites and locations named in the text. The four types of deposit are shown.

### Geology and geomorphology

Vestfold Hills bedrock comprises three dominant suites of Archaean gneisses: Mossel Gneiss, derived from quartzofeldspathic intrusive igneous protoliths; Crooked Lake Gneiss, derived from mafic to felsic potassic intrusive protoliths; and Chelnok Paragneiss, a garnetiferous semi-pelitic metasedimentary suite (Collerson *et al.* 1983, Snape & Harley 1996). The gneisses trend roughly north–south in the far north, and are reoriented roughly NE–SW in the central and southern region (Snape *et al.* 2001). Several suites and sub-suites are structurally repeated on scales of 100s of metres to several kilometres (Snape & Harley 1996), but regionally the orthogneiss protoliths have a range of crystallization ages indicating that structural repetition is less than 10 km (Snape *et al.* 1997). The gneisses are cross-cut by several swarms of Proterozoic mafic dykes that are of variable orientation, but are generally north–south trending and steeply dipping. The dykes range in width from a few millimetres to over 50 m and are areally extensive, forming ~5% of the outcrop (Passchier *et al.* 1991, Dirks *et al.* 1994, Snape *et al.* 2001). Glacial striations reveal former ice flow toward the northwest (Adamson & Pickard 1986a), nearly

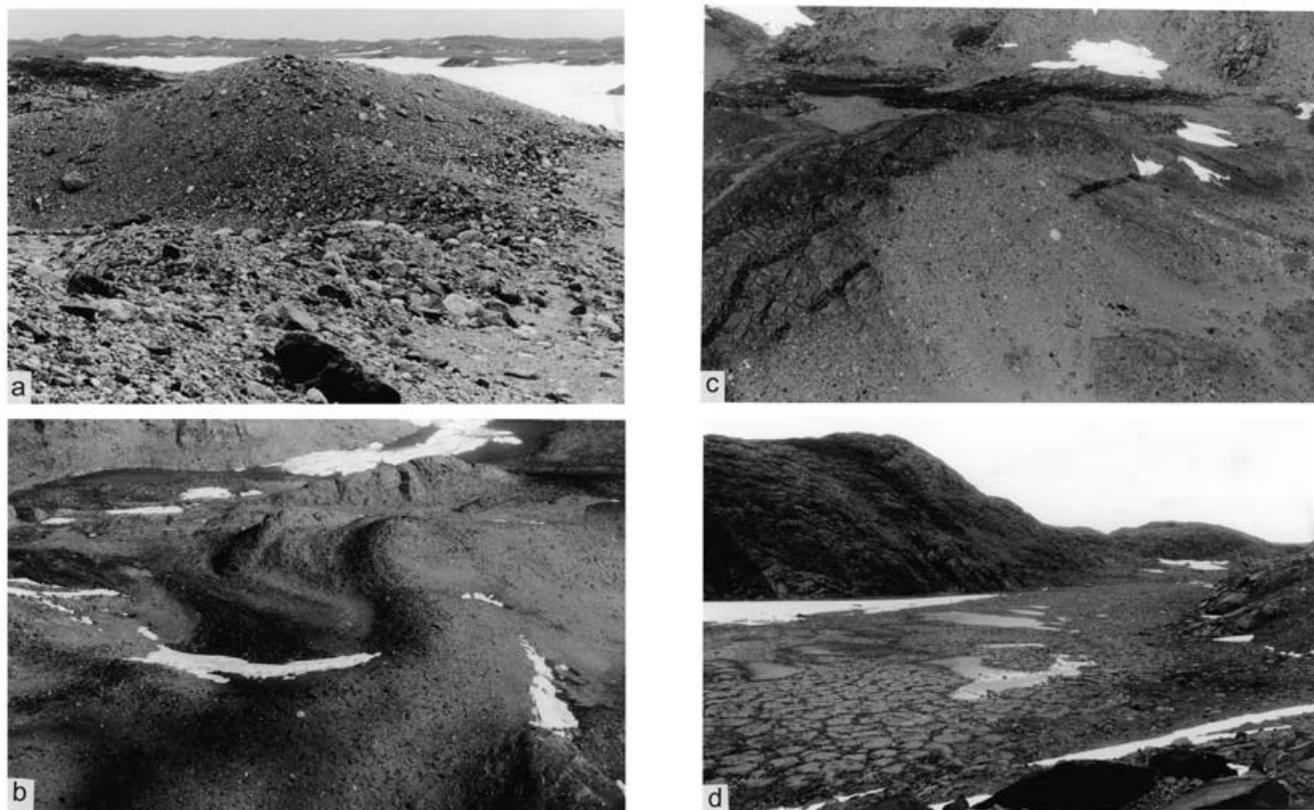
normal to the regional strike of the gneiss and dyke boundaries. Vestfold Hills therefore appears to present a simplified case for detection of sheet-like glacial dispersal patterns.

The glacial depositional landforms in Vestfold Hills were first described by Blandford (1975), who noted the existence of glacier-marginal ridges, and suggested formation via ice push. Adamson & Pickard (1983) mapped the major moraine ridges and noted thick mantles of till. Zhang & Peterson (1984, p. 17) formalized these types of deposits into two morphological groups; ‘irregular surface forms’ (including ridges, hollows and mounds) and till-plain (till-sheet) respectively. They also recognised ‘shear moraine’ (*sensu* Bishop 1957) along the ice sheet margin. Gore (1995) divided the glacial depositional landforms into five morphological types without genetic implication: small debris ridges, large debris ridges, debris drapes, undulating valley fills and flat lying valley fills. Fitzsimons (1997) subsequently divided moraine ridges from Vestfold Hills into types A–D. His type A moraines form where basal debris crops out near the ice sheet margin (*cf.* ‘shear moraines’, *sensu* Bishop 1957). Type B moraines are glaciotectionic features that form from melt out of basal debris concentrated along glaciotectionic fold axes. Type C moraines form as ice contact scree and fans, while type D moraines form as thrust block moraines by freezing-on at the glacier sole, with subsequent transport and in many places glaciotectionic stacking (*sensu* Moran 1971, Moran *et al.* 1980) of till units. Fitzsimons (1990, 1996) also documented the mechanisms and pathways by which glacial sediment remobilises, transfers downslope and where it may ultimately form valley floor sediments. During the initial release of glacial debris from interstitial ice, as well as during sediment deposition, remobilisation and redeposition, there are opportunities for washing and sorting by meltwater to occur. In this way, sediments may become physically and, perhaps, also chemically distinct.

### Methods

#### *Sediment sampling*

In the absence of established transport half-distances for the Antarctic Ice Sheet, an estimate had to be used in order to design the sediment sampling program. Wet based ice sheets move faster and have greater transport half-distances than dry based ice sheets, and there was ample evidence at Vestfold Hills in the form of striations, grooves, chattermarks and crag-and-tail forms (Adamson & Pickard 1983, 1986a, Lundqvist 1989) to indicate that former ice masses there were at least temporarily wet based in most places. In addition, marine fossils and weathered clasts (*i.e.* those exhibiting weathering rinds or tafoni) occur in 62% of



**Fig. 2a.** Small sandy debris ridge on south-eastern Mule Peninsula. The ridge was ~3 m high and was discrete, with adjacent bedrock free of debris. **b.** Large sandy debris ridge on western Broad Peninsula. The ridge was ~20 m high and merged with the surrounding debris surface. **c.** Debris drape on south-eastern Mule Peninsula. This silty diamicton was ~50 m wide, and the disappearance of the mafic dyke shows that the deposit is discrete with the adjacent gneiss bedrock surface completely free of glacial debris. **d.** Cryoturbated valley fill ~100 m across, consisting of silty diamicton, on southeastern Mule Peninsula. Surrounding hillslopes are largely free of glacial debris.

all glacial sediments in Vestfold Hills, and such sediments were most abundant in the western parts of the Hills (Gore *et al.* 1994). Since the marine fossils lay within some kilometres down-ice of the marine fjords, Gore *et al.* (1994) concluded that the glacial debris was deposited within a short distance (< 10 km) of the source. In view of these observations, and as a working estimate in the absence of other data, we estimated a transport half-distance of 5 km, and sampled at an average of one sample per 4 km<sup>2</sup>. We thought this sampling density appropriate for two reasons. First, if the glacial transport distances of the sediments lie at the shorter end of the range suggested by Puranen (1990), say within hundreds of metres of their origin, then the chemical characteristics of the sediment should be similar to the underlying bedrock. Second, if the transport half-distances and thus mixing distances were greater (more than several kilometres), then the sediment chemistry might be poorly related to that of the bedrock and there should be detectable glacial dispersal or 'smearing' across lithological or chemical boundaries. We used the geological map in Collerson & Sheraton (1986) to identify the major suites of Crooked Lake Gneiss, Mossel Gneiss and Chelnok Paragneiss, and sampled sediments from 97 locations across

those boundaries. Sediments were taken from pits that were typically dug to the base of the active layer at 0.3 m to 1.3 m depth. Sub-permafrost excavations, and thus stratigraphy based investigations, were not possible.

Each of the sedimentary deposits was classified into one of four simple landform types. Small debris ridges were typically < 3 m high, 3–4 m wide and 5–20 m long (Fig. 2a) whilst large debris ridges were > 3 m high, 40–50 m wide and up to 1 km long (Fig. 2b). The other deposits readily identifiable were debris drapes (Fig. 2c), which occurred as thin deposits overlying but reflecting the underlying bedrock topography (particularly hillslopes), and valley fill sediments that buried underlying bedrock topography (Fig. 2d). Field sampling occurred during 1989 and 1990, several years prior to Fitzsimons' (1997) four-type depositional model for moraines, and so did not reflect his classification.

#### *Sediment and bedrock geochemistry*

The < 2 mm fraction of 88 of the sediment samples was analysed for major and trace elements using X-ray fluorescence (XRF). Values of ten major elements were

determined with a Philips PW 1404 automatic spectrometer with a dual anode Sc/Mo X-ray tube, at The University of Newcastle (Australia). The matrix was ground in a tungsten carbide ring mill to medium to fine silt size ( $\sim 50 \mu\text{m}$ ), and subsamples were then fused into glass discs for major element analysis, using the method of Norrish & Hutton (1969). An internal monitor of known composition was run daily, and during major element analyses duplicate analyses were made of every fifth sample. Precision of the major element data was assessed using replicate analyses of standards and samples, and is estimated to be better than 1–2% for major elements. Two samples that lay outside the normal range of analytical sums for the major elements (95% to 101%) were considered unreliable analyses and were discarded.

Values of 18 trace elements were determined with a Philips PW 1480 automatic X-ray fluorescence spectrometer with a Rh-anode X-ray tube, at the University of Edinburgh (Scotland). A sub-sample of the ground powder was made into pressed powder discs. Because many analyses had low concentrations of incompatible trace elements, the analytical conditions and calibrations for these elements were optimized for low concentrations where appropriate. Background positions were placed as close as possible to peaks and long count times were used at both peak and background positions. Where background count rates were measured on either side of the peak, as in most trace element determinations, the count time was divided equally between the two positions.

Intensities of trace element lines (La, Ce, Nd, Cu, Ni, Co, Cr, V, Ba, and Sc) were corrected for matrix effects using alpha coefficients based on major element concentrations measured at the same time on the powder samples. Matrix corrections were applied to the intensities of the other trace element lines by using the count rate from the  $\text{RhK}\alpha$  Compton scatter line as an internal standard (Reynolds 1963). Line overlap corrections were applied using synthetic standards.

The spectrometer was calibrated with USGS and CRPG standards, using the values given by Jochum *et al.* (1990) for Nb and Zr, and Govindaraju (1994) for the other elements. Excellent calibration lines were obtained using these standards. Trace element analytical precision was estimated by analysing several standards repeatedly during the analysis of the samples. Some elements were determined at least three times on the low abundance samples, and the results averaged. Reproducibility of most elements is considerably better than 10% relative variation.

The bedrock geochemistry for major and trace elements was taken from Sheraton & Collerson (1984). Analytical techniques were the same as those used for the sediments and operating conditions, and precision and accuracy are similar (or usually better) for the gneisses. The type of bedrock underlying each sediment sample was initially determined from fig. 2.2 in Collerson & Sheraton (1986)

and later confirmed using the latest geological map of the area (Snape *et al.* 2001). Seventy-six geochemical analyses of glacial sediment (those overlying known basement) were grouped according to whether they overlay bedrock Mossel Gneiss, Crooked Lake Gneiss or Chelnok Paragneiss.

#### *Physical analysis of sediment*

The  $< 2 \text{ mm}$  fraction of 97 samples was analysed for grain size (washed at  $63 \mu\text{m}$ , dry sieved at  $0.5 \phi$  intervals from  $2 \text{ mm}$  to  $63 \mu\text{m}$ , X-ray SediGraph 5100 from  $63 \mu\text{m}$  to  $0.5 \mu\text{m}$ ). We calculated sediment mean particle size using the Graphic Mean of Folk (1980), except where the 16th percentile could not be determined, in which case Briggs' (1977) Quartile Mean was applied. Sorting was determined by Graphic Standard Deviation (Folk 1980). Skewness was derived from the Graphic Skewness of Folk (1980), and kurtosis was measured using the Graphic Kurtosis of Folk (1980). Clast macrofabrics were also examined on an opportunity basis, at pits in sixteen (eleven large, five small) ridge sites. At each site, the azimuth and plunge of 30 blade or prolate pebbles or cobbles were recorded from the pit floor. Measurements were plotted on lower hemisphere polar equal area nets, and the data were analysed using the eigenvector method of Mark (1973). Normalised eigenvalues S1, S2 and S3 (Woodcock & Naylor 1983) were calculated to indicate the strength of clustering around their respective eigenvectors.

#### *Data analysis*

Geochemical data were ratioed to remove the constant sum effect (Chayes 1960, 1971) and then  $\log_{10}$  transformed to avoid spurious negative correlations (Aitchison's solution; see Chayes 1949, Aitchison 1984, 1986, Rollinson 1993).

Analysis of the geochemical datasets for glacial carryover patterns was complex. First, the issue of whether or not the bedrock suites were chemically distinct was addressed by testing the null hypothesis that the mean geochemical values of the different basement rock types were the same using one-way analysis of variance (ANOVA; Statistica 5.5). Tukey's pairwise comparison of means (Minitab Release 11) was then used to distinguish which elements made those suites distinct. This was done because if the three bedrock suites were chemically similar, then there would be little reason to expect glacial carryover patterns across the boundaries between them.

Second, the question of whether or not the sediments were chemically distinct from their underlying bedrock was addressed. Multivariate analysis of variance (MANOVA; Statistica 5.5) was used to test the null hypothesis that the mean geochemical values of the different basement rock types and their overlying sediments (grouped according to underlying basement) were the same. The rationale was that if the sediments were chemically indistinguishable from

**Table I.** ANOVA and Tukey results for each of the bedrock suites (data from Sheraton & Collerson 1984).  $P < 0.05$  indicates that there is a significant difference between the bedrock suites; the final column identifies what those significant differences are. MoG = Mossel Gneiss, CLG = Crooked Lake Gneiss, ChP = Chelnok Paragneiss.  $df = 2,74$  for all elements except Ba (2,72).

Element	F	P	Outcomes significant at the 5% level
Major elements			
Si	22.33	0.000	MoG > CLG, ChP
Ti	1.58	0.213	No significant difference between suites
Al	23.45	0.000	MoG > CLG, ChP
Fe	4.14	0.020	ChP > MoG
Mn	0.96	0.387	No significant difference between suites
Mg	5.91	0.004	ChP > MoG
Ca	14.40	0.000	MoG > CLG, ChP
Na	18.20	0.000	MoG > CLG, ChP
K	6.35	0.003	CLG > MoG, ChP
P	21.59	0.000	CLG > ChP > MoG
Trace elements			
Nb	1.07	0.349	No significant difference between suites
Zr	11.74	0.000	MoG > CLG, ChP
Y	4.04	0.022	ChP > MoG
Sr	3.04	0.054	No significant difference between suites
Rb	8.39	0.001	CLG > MoG, ChP
Th	3.80	0.027	No significant difference between suites
Pb	0.01	0.988	No significant difference between suites
Zn	3.57	0.033	MoG > CLG
Cu	2.54	0.086	No significant difference between suites
Ni	3.28	0.043	ChP > MoG
Cr	2.85	0.064	No significant difference between suites
Ce	3.19	0.047	ChP > CLG
La	2.82	0.066	No significant difference between suites
V	4.22	0.019	ChP > MoG
Ba	11.83	0.000	MoG, CLG > ChP

their underlying rock substrate, then that sediment was clearly derived from that rock type and that glacial carryover patterns were not evident.

Third, the relationship between the regional chemistry of bedrock and sediments, and thus the chemical evidence for glacial carryover patterns, was assessed via spatial maps of selected geochemical data.

Fourth, the notion that the sedimentary deposits might be physically distinct was assessed. One-way ANOVA ( $\alpha = 0.05$ ) was used to test the null hypothesis that the mean grain size parameters (mean grain size, sorting, skewness) of the different landform types were the same, and Tukey's pairwise comparison of means to determine which grain size parameters separated the landforms.

Finally, the question of whether or not the sedimentary deposits are chemically distinct was addressed. The sediment geochemical data were examined for significant differences between the four depositional landforms. One-way ANOVA ( $\alpha = 0.05$ ) were used to test the null hypothesis that the mean geochemical values of the different landform types were the same, and then Tukey's pairwise comparison of means was used to determine which elements separated the landforms.

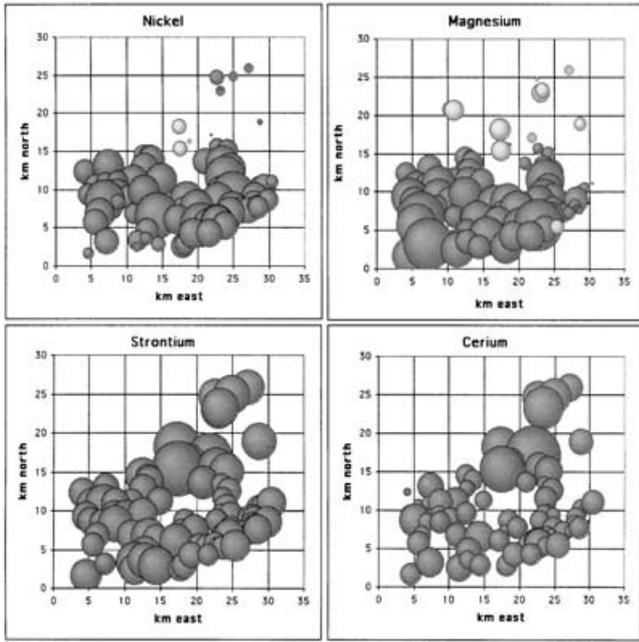
**Table II.** MANOVA P values for the sediment/bedrock data for 24 elements over each of the three major bedrock types.  $P < 0.05$  indicates that there is a significant difference between the sediment and bedrock. Major elements: Mossel Gneiss ( $df = 2,65$ ), Crooked Lake Gneiss ( $df = 2,50$ ) and Chelnok Paragneiss ( $df = 2,35$ ). Trace elements: Mossel Gneiss ( $df = 2,61$ ), Crooked Lake Gneiss ( $df = 2,50$ ) and Chelnok Paragneiss ( $df = 2,32$ ).

Element	Mossel Gneiss	Crooked Lake Gneiss	Chelnok Paragneiss
Major elements			
Si	< 0.001	0.901	0.118
Ti	0.165	0.311	0.356
Al	< 0.001	0.004	< 0.001
Fe	< 0.001	< 0.001	< 0.001
Mn	< 0.001	< 0.001	< 0.001
Mg	< 0.001	< 0.001	< 0.001
Ca	< 0.001	0.037	0.840
Na	< 0.001	< 0.001	< 0.001
K	0.346	< 0.001	0.798
P	0.004	< 0.001	0.084
Trace elements			
Nb	0.076	0.124	0.967
Zr	< 0.001	< 0.001	0.240
Y	0.001	0.114	0.607
Sr	< 0.001	< 0.001	0.001
Rb	< 0.001	0.218	0.019
Th	0.003	< 0.001	0.076
Pb	0.705	0.722	0.928
Zn	0.010	0.689	0.199
Cu	0.001	< 0.001	0.008
Ni	< 0.000	< 0.001	< 0.001
Cr	< 0.000	< 0.001	< 0.001
Ce	< 0.001	< 0.001	0.044
Nd	< 0.001	< 0.001	0.359
La	< 0.001	< 0.001	0.003
V	< 0.001	0.001	0.957
Ba	0.339	< 0.001	< 0.001
Sc	0.742	< 0.001	< 0.001

## Results

### *Glacial carryover of sediment*

If the bedrock suites were chemically indistinguishable, then no glacial carry over patterns would be expected across their boundaries. Of the 26 elements studied, only nine were statistically indistinguishable throughout the three bedrock suites (Table I). Mossel Gneiss was the most dissimilar of the three suites and is furthest in composition from the Chelnok Paragneiss, which is consistent with the mineral assemblages, protolith character, and geochemical changes associated with metamorphism. The Mossel Gneiss, for example, had higher Si and Al (reflecting a greater abundance of quartz and plagioclase feldspar) and lower Fe, Mg and Ni (reflecting less hornblende, garnet and biotite) than the Chelnok Paragneiss (Collerson *et al.* 1983, Snape & Harley 1996). Similarly, both the Mossel Gneiss and the Chelnok Paragneiss had lower K and Rb than the Crooked Lake Gneiss, a feature that is attributed to the loss of large ion lithophile elements during the peak high grade

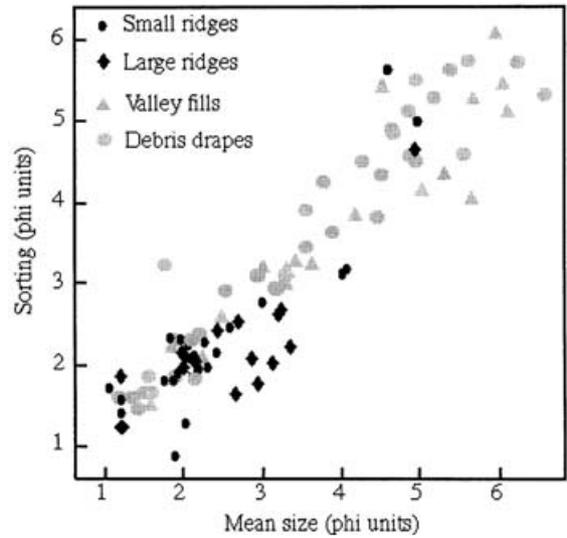


**Fig. 3.** Spatial maps of selected sediment chemistry using transformed data. Dark circles are values > 0, clear circles are values < 0. Circle width is proportional to the magnitude of the value (raw and transformed data are at [www.antdiv.gov.au/aadc](http://www.antdiv.gov.au/aadc)). The grid has an arbitrary origin at the westernmost end of Mule Peninsula.

metamorphic event that preceded the intrusion of the Crooked Lake protoliths (Collerson & Sheraton, 1986, Snape *et al.* 1997). It is also not surprising to find that the Crooked Lake Gneiss overlapped with the other two suites to some extent because it is compositionally very diverse, with rock types ranging from ultramafic cumulates, to tonalites and quartz monzonites that are chemically very similar to the Mossel Gneiss.

The sediments were largely dissimilar to their underlying bedrock. The sediments above Mossel Gneiss were dissimilar to the underlying bedrock in 21 of the 27 elements, the sediments above Crooked Lake Gneiss were dissimilar in 20 of the 27 elements, and Chelnok Paragneiss was dissimilar in 14 of the 27 elements (Table II).

There is also chemical evidence of glacial carryover patterns. There is a distinct pattern of regional sediment chemistry that is caused by three factors. First, there is the influence of marine salts. The area below 10 m a.s.l., the



**Fig. 4.** Bivariate plot of mean grain size versus sorting, using untransformed data. Both small and large debris ridges are clustered at the lower left of the plot, indicating larger grain size and better sorting while debris drapes and valley fills are equally represented along the distribution of both grain size and sorting.

height of the Holocene marine incursion, as well as the area west of the ‘salt line’ (*sensu* Blandford 1975, Adamson & Pickard 1986b, Gore *et al.* 1996), is saturated with marine salts that have changed sediment bulk chemistry. The salt line effect was very clearly expressed in the sediment chemistry, and we removed from further consideration those elements, such as Na and Mg, that reflected the input of salt spray. Second, the sediments are also greatly enriched in P north of the ‘Death Valley’ (*sensu* Korotkevich 1971) system of lakes on Broad Peninsula and on Tryne Peninsula, with the greatest concentrations near to penguin rookeries at Anchorage Island and Rookery Lake. These cases are believed to reflect the input of penguin excrement, rather than anomalous concentrations of minerals such as apatite. Third, after removal of the marine salts and P, a subtle residual pattern emerged that we ascribe to glacial carryover. The sediments north of the Death Valley system of lakes to the northern coast of Tryne Peninsula are deficient in Ni and Mg relative to the southern sediments, while the same sediments exhibit enhanced concentrations of Sr and Ce. (Fig. 3). This pattern of sediment chemistry is consistent with the area of outcrop of metasediments on Tryne Peninsula and the northern edge of

**Table III.** ANOVA and Tukey results for four grain size parameters; the final column identifies what those significant differences are.  $P < 0.05$  indicates that there is a significant difference between the deposits. SDR = small debris ridges, LDR = large debris ridges, DD = debris drapes, VF = valley fills.

Grain size parameter	F	Df	P	Outcomes significant at the 5% level
mean	5.21	3,93	0.002	SDR are coarser than DD and VF
sorting	7.19	3,90	0.000	SDR and LDR are better sorted than DD and VF
skewness	6.00	3,90	0.001	SDR are more positively skewed than DD and VF
kurtosis	2.13	3,39	0.113	No significant difference between deposits

**Table IV.** ANOVA and Tukey results for 10 major elements and 18 trace elements.  $df = 3,83$  for major elements and  $df = 3,77$  for trace elements.  $P < 0.05$  indicates that there is a significant difference between the deposits. The fourth column identifies what those significant differences are. SDR = small debris ridges, LDR = large debris ridges, DD = debris drapes, VF = valley fills. The last column identifies minerals for which the corresponding element has a strong affinity, based on partition coefficients in Rollinson (1993, p. 108–111).

Major elements	F	P	Outcomes significant at the 5% level	Associated minerals
Si	4.07	0.010	SDR > DD	Quartz
Ti	2.67	0.053	No significant difference between deposits	
Al	0.57	0.635	No significant difference between deposits	
Fe	1.75	0.164	No significant difference between deposits	
Mn	1.51	0.217	No significant difference between deposits	
Mg	2.22	0.093	No significant difference between deposits	
Ca	2.67	0.053	No significant difference between deposits	
Na	1.63	0.189	No significant difference between deposits	
K	3.24	0.027	DD > SDR	K-feldspar
P	2.49	0.067	No significant difference between deposits	
Trace elements				
Nb	5.84	0.001	DD and VF > SDR	Biotite, sphene, hornblende
Zr	5.41	0.002	SDR > VF, LDR > DD & VF	Hornblende
Y	6.33	0.001	SDR > DD	Apatite, garnet, hornblende, clinopyroxene
Sr	0.44	0.722	No significant difference between deposits	
Rb	3.11	0.031	DD > SDR	Biotite
Th	1.15	0.335	No significant difference between deposits	
Pb	1.44	0.237	No significant difference between deposits	
Zn	3.83	0.013	VF > SDR	No data
Cu	2.27	0.087	No significant difference between deposits	
Ni	1.58	0.201	No significant difference between deposits	
Cr	1.64	0.187	No significant difference between deposits	
Ce	1.79	0.156	No significant difference between deposits	
Nd	1.20	0.315	No significant difference between deposits	
La	1.83	0.149	No significant difference between deposits	
V	1.58	0.201	No significant difference between deposits	
Ba	0.69	0.561	No significant difference between deposits	
Sc	3.60	0.017	SDR > DD	Hornblende, garnet, magnetite, zircon, biotite, ilmenite

Broad Peninsula. The distance over which the transition occurs is  $< 5$  km, and probably  $< 3$  km. The distance of this transition is interpreted to be a distance over which the former chemical character of the sediment has been lost. That is, the glacial transport half-distance is probably  $< 3$  km.

#### *Attributes of the glacial landforms*

The sedimentary deposits were found to be both physically and chemically distinct. Small debris ridges were different from debris drapes and valley fills in terms of mean grain size, sorting and skewness (Table III). In addition, large debris ridges were different from debris drapes and valley fills in terms of sorting. A general trend emerged that the debris ridges, particularly the small ridges, are coarser grained and better sorted than the debris drapes and valley fills (Fig. 4). Sediments can become coarser grained as fine grained, lighter and platy minerals are winnowed from the sediment by meltwater and washed away. Such minerals particularly include micas and clays. These textural changes can lead to chemical differentiation of the deposits. The small debris ridges were distinct from debris drapes for three major elements, but with no difference for seven of the

major elements (Table IV). Small debris ridges were distinct from the debris drapes and valley fills for six trace elements, but with no significant difference for 12 of the trace elements.

#### **Discussion**

The first aim of this study was to determine whether or not the debris of Vestfold Hills was from local or distant sources. The bedrock suites, particularly the Mossel Gneiss, are distinct enough in bulk chemistry to justify examining the sediments for glacial carryover patterns across the boundaries of those suites. The sediments were distinct from their underlying bedrock, with an apparent increase in similarity between the rocks and sediments of Mossel Gneiss, Crooked Lake Gneiss and Chelnok Paragneiss, in that order. There are several possible explanations for this, the most likely being the relative ease with which the different basement types are eroded by glacial action. Generally the gneisses become progressively more schistose in the order Mossel Gneiss – Crooked Lake Gneiss – Chelnok Paragneiss, which in turn largely reflects the abundance of platy minerals such as biotite and hornblende, and overall outcrop scale homogeneity.

Apart from glacial carryover between lithologies, there are two other reasons why the sediment geochemistry may be dissimilar to that of the underlying basement. First, there may have been mineralogical, and thus chemical differentiation of sediments during deposition. The mechanism envisaged is fluvial winnowing of smaller or lighter minerals during release of the sediment from glacial ice. This idea was examined below, through an analysis of the grain size characteristics of the four landform types. Second, glacial erosion of mafic dykes may contribute a minor, but significant component to the sediment. In places, cobble and boulder trains lie down-ice of the prominent dykes that have an average spacing of ~0.1 km across Vestfold Hills. In the field, those trains typically overlapped other dykes and their trains, so the distance of glacial transport of these clasts could not be determined. However, there was abundant physical evidence in the form of these trains and finer lithic fragments, for glacial carryover of least 0.1 km. There is also no doubt of the presence of a long-travelled, but minor component of the till. Occasional conspicuous erratics of unknown provenance were found across the Hills, including a distinctive banded iron stone and an omphacite-bearing gneiss, that testify to the long-distance (> 20 km) transport of some clasts. However, the geochemical data show that the bulk of the sediment at Vestfold Hills has been derived locally, consistent with observations of marine fossils and weathered clasts in the glacial sediment (Gore *et al.* 1994). Consideration of these results and the spatial patterns of bedrock chemistry, allowed us to conclude that the sediment transport half-distance was less than 5 km but greater than 0.1 km. The most likely distance is < 3 km (Fig. 3), which is at the shorter end of the range reported by Puranen (1990) for the Fennoscandian ice sheet.

A short transport distance means that most of the sediment in the western parts of the Hills will have experienced several subsequent erosional and depositional events before being ultimately carried onto the continental shelf by the ice sheet, and that there must have been multiple opportunities for fluvial winnowing during deglaciation, and physical and chemical weathering during interglacial periods. Gore & Colhoun (1997, figs 2 & 3) noted that the sediments in western parts of Vestfold Hills are coarser and better sorted than those in the eastern parts. They suggested that this was probably due to the addition of coarse, lithic sands from salt weathering of bedrock and clasts during interglacial periods. However, multiple cycles of fluvial winnowing is also consistent with this spatial pattern of sediment grain size. It is likely that these two processes act in concert.

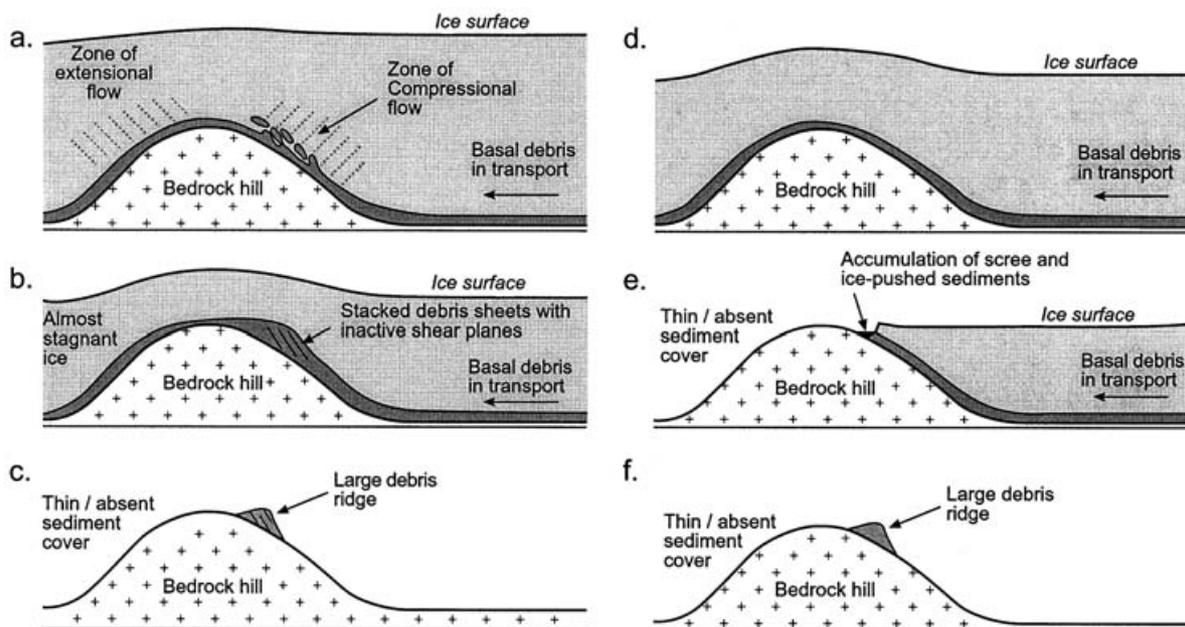
The second aim of this study was to determine whether or not the sediments in those landforms were physically and chemically distinct from each other. We defined four landforms and Table III shows that the small debris ridges are coarser, better sorted and more positively skewed than

the debris drapes and valley fills. In addition, the large debris ridges are better sorted than the debris drapes and valley fills. These ridges, and particularly the small ridges, reflect marked fluvial sorting during deposition.

This idea was supported by the deposit geochemistry. Table IV shows that the small debris ridges are depleted in K, Nb, Rb and Zn, and enriched in Si, Zr, Y and Sc. Partition coefficients reflect the propensity for elements to become incorporated in particular minerals (Rollinson 1993). These coefficients can be used to predict that the small ridge sediments are depleted in a mineral assemblage that includes biotite and K-feldspars, while being enriched in a mineral assemblage that includes quartz, hornblende, garnet, clinopyroxene and orthopyroxene.

Examination of the sediments from the debris drapes revealed that the fine grained fraction was typically rich in biotite and easily comminuted feldspars. Examination of the coarse grained debris ridge sediments revealed that they were depleted in these minerals relative to the mineral assemblage in the debris drapes. The characteristics that control the behaviour of these minerals - some depleted, some enriched - are those that control their hydraulic behaviour, *viz.* grain size, form and specific gravity. Minerals depleted in the ridge sediments are small, platy and have a relatively low specific gravity, and the minerals enriched in the ridge sediments tend to be large, not platy and have a relatively high specific gravity. The ridge sediments appear to have had their mineral assemblages, and thus their bulk chemical composition, modified as a consequence of fluvial winnowing during sedimentation. The small debris ridges have been modified to a much greater extent than the large debris ridges. In contrast, the debris drapes and valley fills appear to have undergone little if any modification by fluvial processes during deposition.

Fitzsimons (1997) identified four types of moraine ridges, A–D. The small debris ridges in this study equate with Fitzsimons' type A, B or C moraines. His type A ridges are 'shear' (*sensu* Bishop 1957) or 'inner' (*sensu* Weertman 1961) moraines, that form as ice cored mounds in a supraglacial position near the ice sheet margin. His type B ridges form from debris concentrations along subglacial recumbent fold axes. His type C ridges are ice contact scree and fans which form at ice cliff margins. Fitzsimons (1991) also identified supraglacial eskers that produce ridges of well washed sediments. When the interstitial ice has been removed fully from Fitzsimons' (1997) A, B and C moraine ridges and (1991) eskers, and the sediments have been 'let down' to their final position, a small debris ridge consisting of well washed sands will result. The inferred fluvial winnowing leading to the formation of our small debris ridges is thus consistent with Fitzsimons' (1997) model for the genesis of ridge types A–C, as well as his eskers (Fitzsimons 1991). While it is clear from contemporary field observations that the type A–C moraines and the supraglacial eskers have different genetic pathways



**Fig. 5.** Glaciotectonic stacking of till sheets (after Moran 1971) to form Large Debris Ridges (type D moraines) on the sides of hills. **a.** During inundation by ice, basal debris is in transport across the rock surface. Compressional flow on the stoss (up-ice) side of hills causes shearing of till sheets and stacking of debris bands. **b.** As ice flow slows, the stacked debris sheets melt out *in situ*, creating **c.** a Large Debris Ridge with internal thrust geometries. Stationary ice margins near hilltops allow the accumulation of Large Debris Ridges on the stoss side of the hill. **d.** During inundation by ice basal debris is in transport across the rock surface. **e.** Upon thinning and retreat, the ice ‘pins’ against hillsides for long periods, allowing the accumulation of ice-contact scree or the emplacement of thrust block moraines. **f.** Removal of the ice leaves a Large Debris Ridge on the stoss side of the hilltop.

(Fitzsimons 1991, 1997), it is not clear to us that they will be readily distinguishable away from the ice margin, with the possible exception of the type C ridges which may have a stronger fabric, or the eskers that might be distinguishable because they would form orthogonal to the former ice margin.

We took fabric data from Gore (1995, Appendix 1; one type A ridge, four type C ridges) and combining them with those from Fitzsimons (1997; four type A ridges, eight type B ridges, 15 type C ridges), we examined whether or not fabric alone could distinguish the moraine ridges type A–C. One-way ANOVA on the normalised eigenvalue parameter S1 (*sensu* Mark 1973), showed that there were no significant differences between Fitzsimons’ (1997) type A–C ridges, either using his data alone ( $F(2,24) = 1.80$ ;  $P < 0.187$ ) or our combined data ( $F(2,29) = 1.64$ ;  $P < 0.212$ ). Similarly, there were no significant differences between the ridge types in terms of the S3 parameter, either using his data alone ( $F(2,24) = 1.08$ ;  $P < 0.355$ ) or our combined data ( $F(2,29) = 1.19$ ;  $P < 0.320$ ). Since the type A–C ridges are similar in size, morphology, sediment grain size and clast fabric, then they are essentially indistinguishable in the sedimentary record, away from the ice margin where they formed. We conclude that until diagnostic criteria for their distinction are better defined, the moraine ridge types A–C

of Fitzsimons (1997) should be collectively regarded as ‘small debris ridges’ as defined in this study. In the field, their small deposit size, coarser grain size, better sorting and positive skewness of the sediments, and the resultant effects on the geochemistry, are typical of the small debris ridges.

The large debris ridges in this study equate directly with Fitzsimons’ type D moraines, which he describes as thrust block moraines that form when unconsolidated sediment is entrained by freezing, shearing and thrusting of sediment blocks. The large moraine ridges are sandy, commonly contain marine fossils, usually lie down-ice of marine inlets, but typically lie on the up-ice side of hills or at hilltops. These field relationships and the physical and chemical characteristics of the sediments indicate that the deposits probably form either via glaciotectonic stacking of subglacial till sheets in the way that Moran (1971) and Moran *et al.* (1980) envisaged, or as thrust block moraines as Fitzsimons (1997) suggests. Under Moran’s model, compressional flow of ice led to stacking of debris subglacially on the stoss side of the hilltops, while extensional flow of ice on the leeward side of the hill resulted in few moraines forming there (Fig. 5a). Alternatively, the moraines may mark the former position of an ice edge, with the moraine forming through bulldozing, thrust block emplacement or as an ice contact scree

(Fig. 5b).

Whatever mechanism is invoked has to explain the rather large amounts of debris that form these deposits. In either of these mechanisms, ice having crossed the water saturated, debris rich floors of marine inlets would erode and transport sufficient material to form these large debris ridges. It is no coincidence that in Vestfold Hills, the occurrence of large debris ridges is restricted to the down-ice side of marine inlets.

Table III shows that the large debris ridges are transitional between the small debris ridges (type A–C moraines), debris drapes and valley fills, in terms of mean grain size, sorting and skewness. Similarly, Table IV shows that the large debris ridges are transitional in terms of their chemistry. We then examined whether or not the type D ridges have distinctive fabrics. We took fabric data from Gore (1995, Appendix 1; one type A ridge, four type C ridges, 11 type D ridges) and combining them with those from Fitzsimons (1997; four type A ridges, eight type B ridges, 15 type C ridges, six type D ridges), we examined whether or not fabric alone could distinguish the moraine ridges types A–D. Unfortunately, their fabrics are not distinctive. One-way ANOVA on the normalised eigenvalue parameter S1 (*sensu* Mark 1973), showed that there were no significant differences between Fitzsimons' (1997) ridge types A–D, either using his data alone ( $F(3,29) = 1.60$ ;  $P < 0.210$ ) or our combined data ( $F(3,45) = 1.75$ ;  $P = 0.170$ ). Similarly, there were no significant differences between the four ridge types in terms of the S3 parameter, either using his data alone ( $F(3,29) = 1.24$ ;  $P < 0.312$ ) or our combined data ( $F(3,45) = 0.90$ ;  $P < 0.449$ ). Since the large debris ridges (type D moraines) are not readily characterised in terms of grain size, clast fabric or geochemical parameters, the best way that they can be identified is either via their large size, or, more reliably, via recognition of shear planes in their stratigraphy.

The remaining two landform types that we defined, debris drapes and valley fills, are similar in terms of their physical and chemical characteristics. Fitzsimons (1990, 1996) described how debris redistributes rapidly and effectively following deglaciation, leading to concentrations of paraglacial sediment on hillslopes and in valley fills. There is no physical or chemical evidence for sorting of any kind during the lateral transfer from hillslope to valley bottom that Fitzsimons (1990, 1996) describes, although it is possible that valley fills may acquire the fabrics of mass movement deposits. We now believe that our debris drapes and valley fills are two morphological manifestations of Fitzsimons' (1996) paraglacial sediments, and that our field-based distinction did not assist our understanding of the genesis of those deposits.

## Summary

Sediment geochemistry, weathered clasts and marine fossils

indicate that the bulk of the sediment at Vestfold Hills is derived locally, and deposited within 5 km and probably within 3 km of its source. This short transport distance means that there were multiple opportunities for subsequent fluvial winnowing of some sediments during deglaciation, and for physical and chemical weathering during interglacial periods. We defined glacial deposits with four morphologies; small debris ridges, large debris ridges, debris drapes and valley fills. There is a progression in fluvial sorting during deposition from debris drapes and valley fills (least washed), to large debris ridges (better washed) to small debris ridges (most washed). As a consequence, the small ridges were the most distinct in terms of grain size parameters and chemistry. The type A–C moraines of Fitzsimons (1997) were not readily distinguished in terms of morphology, grain size or clast fabric and we refer to them collectively as small debris ridges. The large debris ridges are transitional in grain size and chemistry between the small ridges and the debris drapes and valley fills. The debris drapes and valley fills were not distinguishable in terms of grain size parameters or chemistry, and distinctions between them in terms of sedimentology are not meaningful. We conclude that they are two morphological manifestations of the paraglacial sediments of Fitzsimons (1996). An implication of the short transport distance and the widespread occurrence of large debris ridges – interpreted to be thrust block (type D) moraines – is that the stratigraphy of the glacial deposits should be expected to be complex, with an admixture of sediments and fossils of various ages in any stratigraphical section.

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