NORSK POLARINSTITUTT
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Sedimentation of the Wood Bay Formation (Devonian) of Spitsbergen: Regional analysis of a late orogenic basin

NORSK POLARINSTITUTT
OSLO 1972

DET KONGELIGE DEPARTEMENT FOR INDUSTRI OG HÅNDVERK

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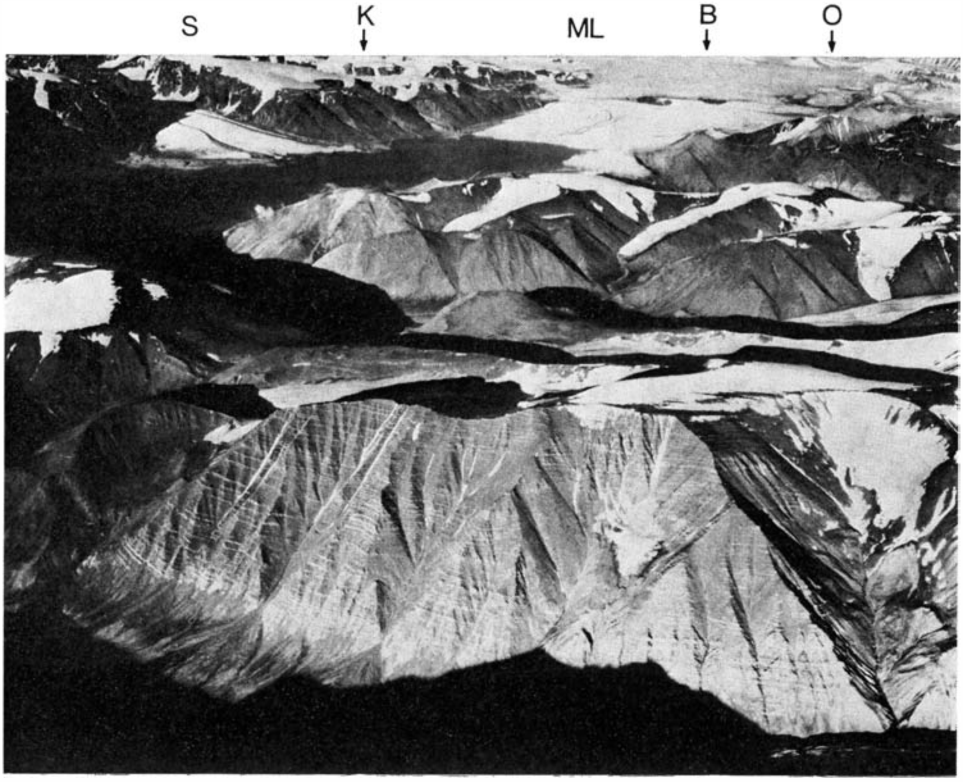

WOOD BAY FORMATION IN EASTERN DICKSON LAND

In the left background, the Hecla Hoek (pre-Devonian) mountains of Ny Friesland round Stubendorffbreen (S) lie behind Austfjorden.

The boundary fault (Balliolbreen fault) runs in a north-south direction beneath Austfjorden, and then outcrops on mountains such as Odellfjellet (O), on this side of Mittag-Lefflerbreen (ML).

On this side of the boundary fault, light-coloured outcrops in Kastellet (K), Bulmanfjellet (B), and Odellfjellet (O) are exposures of the Austfjorden Sandstone Member of the Wood Bay Formation. They were deposited by large braided rivers draining an extensive area of high relief to the south-east.

In the foreground, the face of Sir Thomasfjellet shows the cyclic deposition which is characteristic of most of the rest of the Wood Bay Formation. Resistant sandstone units alternate with less-resistant red (darker) siltstone units. The sandstones were deposited by the channels of rivers which flowed generally from the south. The red siltstone deposits accumulated in the interchannel areas of the floodplain. The increase in the proportion of sandstone towards the lower part of the mountainside probably reflects a higher proportion of sand bed-load in the rivers.

Contents

Abstract

Our object has been the analysis of the fresh-water, late orogenic, Wood Bay Formation, which outcrops over an area 150 km by 75 km, and is up to 3 km thick.

Expeditions from Cambridge University have collected systematic sedimentological data, mainly measuring "vertical" sections. Vertebrate fossils were used to define three informal time-rock subdivisions of the Formation.

Analysis of palaeocurrents, sandstone composition and grain-size variation have allowed us to distinguish three river systems which flowed towards a northern area of clay flats. Rivers of the eastern system were large, north and north-north-west flowing, bed-load rivers of braided type. They were relatively heavily laden with sand-grade sediment, which was rich in feldspar. In contrast, the rivers of the western system were small, eastward flowing, mixed-or-suspended load rivers of high sinuosity. They were relatively less loaded with sand-grade detritus, which was poor in feldspar, particularly in "Lower" Wood Bay time. The central system flowed north, but was similar to the western system in other respects.

We suggest that the eastern system drained an area to the south-east that was very large compared with the small drainage areas of the western system. Calculation of the relative size of these areas depends on estimation of the relative effective precipitation of rain. If this relationship is estimated, and an actual area worked out for a western river, areas, denudation rates and relief can be suggested for all the systems. Suggestions for the relief are 150 m in the west, and 6000 m in the south-east. Estimates of vertical movement can be made using these figures and a simplified model of tectonism followed by isostatic adjustment, erosion and sedimentation.

Fold trends provide evidence that the Upper Devonian eastern boundary fault (Balliolbreen Fault) is a left-lateral, strike-slip fault. The isolated Devonian succession round Hornsund can most easily be fitted to the regional pattern in the north, by supposing that the Hornsund rocks have been moved by major left-lateral strike-slip faulting too.

Introduction

TECTONIC SETTING AND OBJECTS OF OUR WORK

The Wood Bay Formation is the most extensive formation in the late Silurian and Devonian succession of Spitsbergen. This late Silurian and Devonian succession consists of fresh or brackish-water deposits of great thickness (8 km is the sum of the greatest thicknesses of the various formations). It accumulated after major Caledonian folding and metamorphism, and was then deformed by an Upper Devonian (Svalbardian, VOGT 1929) phase of folding and faulting. Because of these tectonic features, this depositional episode is regarded as a late part of the Caledonian orogeny (HARLAND 1961, Fig. 5).

The Wood Bay Formation occurs predominantly in Northern Spitsbergen (Fig. 1), where it outcrops over an area 150 km by 75 km, and is up to 3 km thick.

Our object in this work has been to analyse this sheet of sediment to determine the processes responsible for it.

In this paper, we firstly describe our technique for sampling and recording the

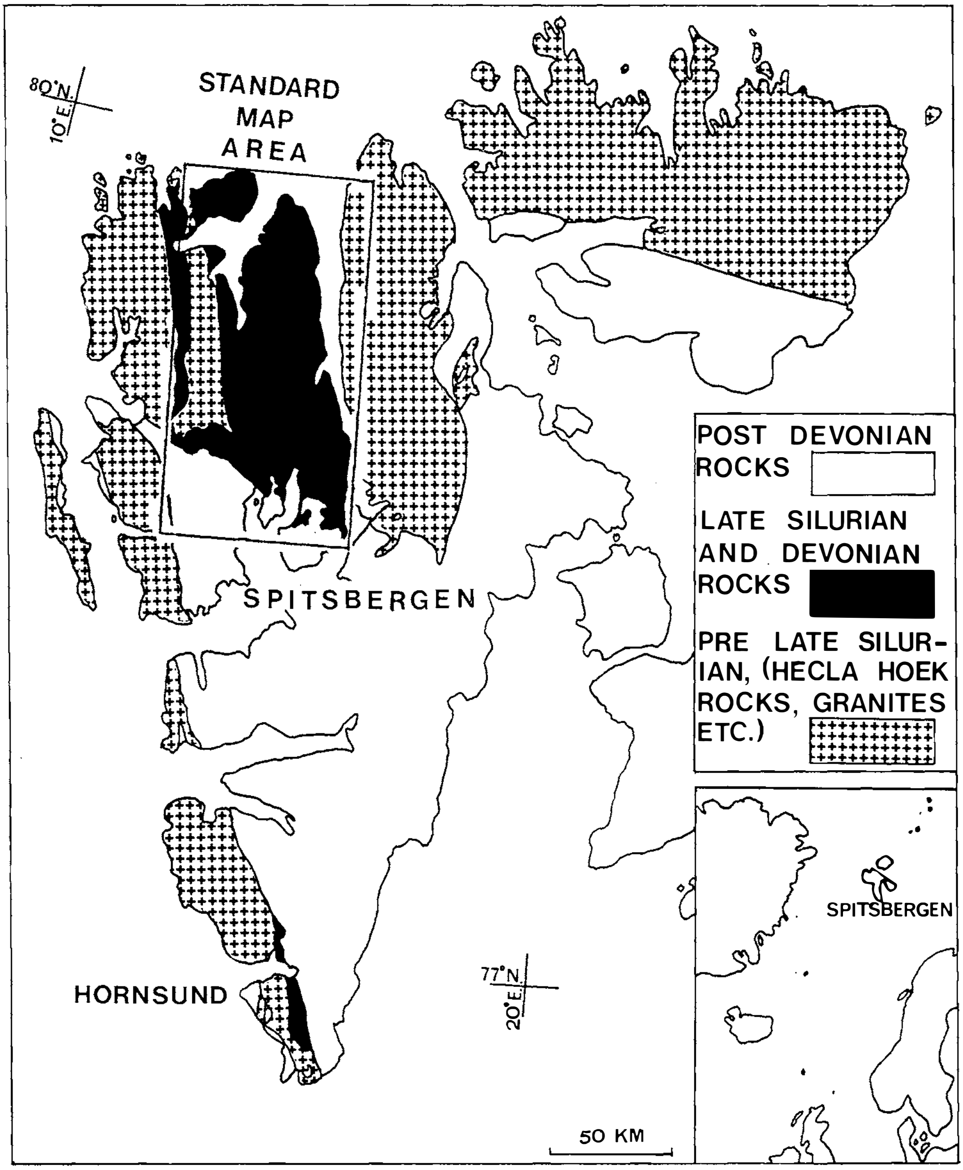

Fig. 1. *Map of Spitsbergen, showing the outcrops of the late Silurian and Devonian succession, and the limits of the 'standard map area' used in much of this paper.*

local details of the Wood Bay Formation. We then describe the regional analysis of the river palaeocurrents, the sandstone composition and the grain-size variation.

We have used laboratory studies of water flow, and knowledge of present-day rivers, to interpret the features of the Wood Bay Formation in terms of deposition on clay-flats and by rivers of characteristic channel type, depth, flow strength and slope. We have then explained these systems in terms of differences of water and sediment discharge. Finally our analysis has suggested climatic and tectonic controls, such as rainfall and source area size, relief and uplift.

TOPOGRAPHY AND ORGANISATION OF FIELD-WORK

The areas (Figs. 1, 2, 5) in which the Wood Bay Formation outcrops are mountainous, with heights ranging from 400 m in the north to 1500 m further south. These mountains form a relatively ice-free belt, lying between the ice-sheets of Ny Friesland and James I Land. All the place names we have used are listed in "Place-Names of Svalbard" (Norsk Polarinstitutt Skrifter Nr. 80 and 112, 1944 and 1958), or appear on the latest maps published by Norsk Polarinstitutt.

Our field-work was carried out by Friend on seven summer expeditions and Moody-Stuart on four summer expeditions, between 1955 and 1965. These expeditions were part of the programme of geological investigation in Spitsbergen initiated and directed by W. B. Harland of the University of Cambridge. He pointed out the interest of these rocks to us, and gave us his observations made in Andrée Land in 1951, and Dickson Land in 1953. He also made us think about the tectonic significance of our work.

Below we give a list of the expeditions in which we were directly involved, and refer to the published report of each expedition. We should also like to take this opportunity of thanking all those who were members of our field parties.

1955 (Lobban 1956): B. Moore
1957 (Friend 1958): M. J. Allderidge, M. G. Bawden, P. T. Warren
1958 (Friend 1959): R. F. Atherton, D. B. Benton, M. D. Fuller, D. J. Gobbet, C. J. B. Kirton, J. C. Rucklidge, P. R. Simpson, J. C. Taylor
1959 (Harland 1960): R. A. Gayer, D. G. Gee, D. W. Matthews
1961 (Friend 1962): K. C. Allen, M. J. Collins
1962 (Harland 1963): I. E. Scholey, A. H. Neilson
1963 (Harland 1964): A. J. Wainwright, M. C. Barr, A. Jenkinson
1964 (Harland 1965a): A. Maynard-Smith, D. J. W. Piper
1965 (Harland and Wallis 1966): J. C. Moody-Stuart, R. Ferguson, P. Coopman, I. A. D. Sweetman.

Professor O. M. B. Bulman and the staff of the Department of Geology, University of Cambridge, provided the home support for these expeditions. Mrs. K. N. Herod helped with numerous jobs, particularly with the integration of our records with Mr. Harland's Spitsbergen filing system.

Many Norwegians have helped with advice and support of all sorts. We should like particularly to thank Professor Anatol Heintz and Dr. Natascha Heintz, and also Dr. Tore Gjelsvik and his staff at the Norsk Polarinstitutt.

The cost of our field work has been partly borne by contributions from all the members participating. Major grants came from the British Department of Scientific and Industrial Research. Other grants came from the Royal Geographical Society, Gino Watkins Memorial Fund, Mount Everest Foundation, Shell International Petroleum Company Limited. Equipment and stores were provided at favourable rates by many British firms.

Friend received a maintenance grant from the British Department of Scientific

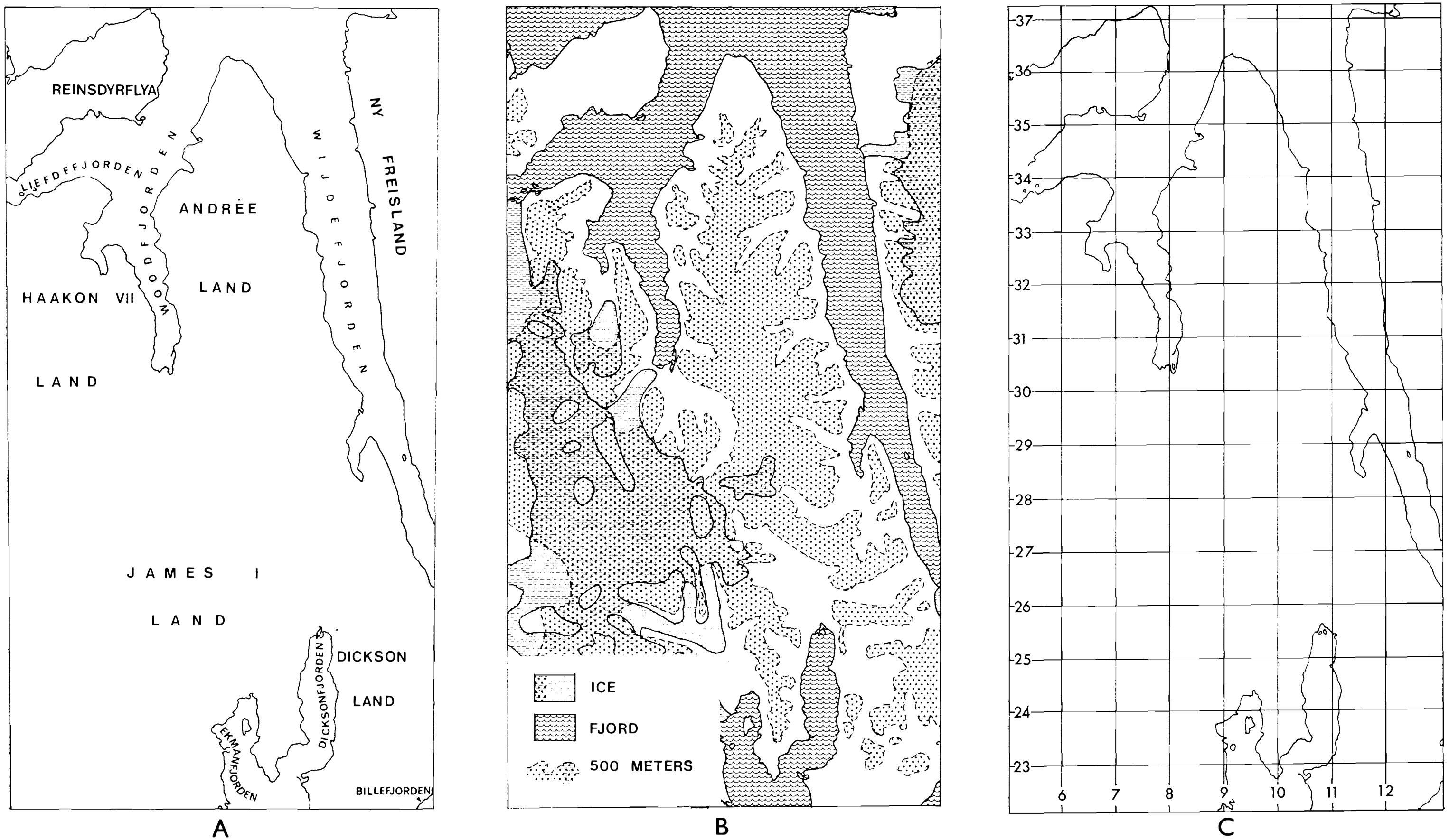

Fig. 2. *The standard map area (defined in Fig. 1), showing a) main place names, b) topography and ice-sheet distribution, c) standard grid.*

and Industrial Research and was, later, a Research Fellow of Gonville and Caius College, Cambridge. MOODY-STUART received a maintenance award from Shell International Petroleum Company.

METHODS

Sedimentological Sections

Our earlier expeditions were devoted to topographical (HARLAND and MASSON-SMITH 1962), stratigraphical (FRIEND 1961) and structural mapping of the Devonian terrain, generally on a scale of 1/50,000. It was only at a later stage that we came to realise the potential for detailed sedimentary analysis. Some sedimentological features of the Wood Bay Formation have previously been described by BIRKENMAJER (1965), DINELEY (1960), FRIEND (1961, 1965), FRIEND and MOODY-STUART (1970), MOODY-STUART (1966).

In our sedimentary analysis, our main field technique has been the measurement of "vertical" sedimentological sections. These sections have varied from 10 m to 470 m in length (Appendix 3). They were usually measured by an observer and a recorder working their way up a gully or ridge. After some experimentation, we settled on the use of a printed form (Fig. 3), small enough to be folded into a field notebook. The recorder prepared a graphical plot of the section, set by set, while it was being investigated in the field. Further discussion of the measurement of "vertical" sedimentological sections (along with a later development of the recording technique) has been presented by ALEXANDER-MARRACK, FRIEND and YEATS (1971).

All of these sections have been located in the field on our draft 1/50,000 maps. We have then used the Norsk Polarinstitutt grid system to calculate coordinates for each section, and these have been used in our analytical maps. Most of the maps in this paper refer to a standard part (50–130 E., 222–382 N.) of this coordinate grid (Figs. 1, 2).

Analysis of data

The analysis reported in this paper, has been made over several years. From 1966 this work was carried out by FRIEND as a member of the staff at the Scott Polar Research Institute, University of Cambridge. We would like to thank Dr. G. DE Q. ROBIN and the other staff of the Institute, for their support in this work. Miss ANNE SWITHINBANK has particularly helped with the preparation of this paper. We should also like to thank the friends who have read this paper at various stages.

We have made use of the University of Cambridge Titan Computer in our analytical work. All our section data on grain-size and thickness of sets have been punched onto paper tape, and analysed by a number of simple programs.

We have also made use of a standard display method in our analysis of the variation of certain parameters across our area. This method is known as iterative-fit trend-surface analysis (COLE 1968, 1969; READ, DEAN and COLE 1971). The actual program we used was written by B. M. E. SMITH of the Scott Polar

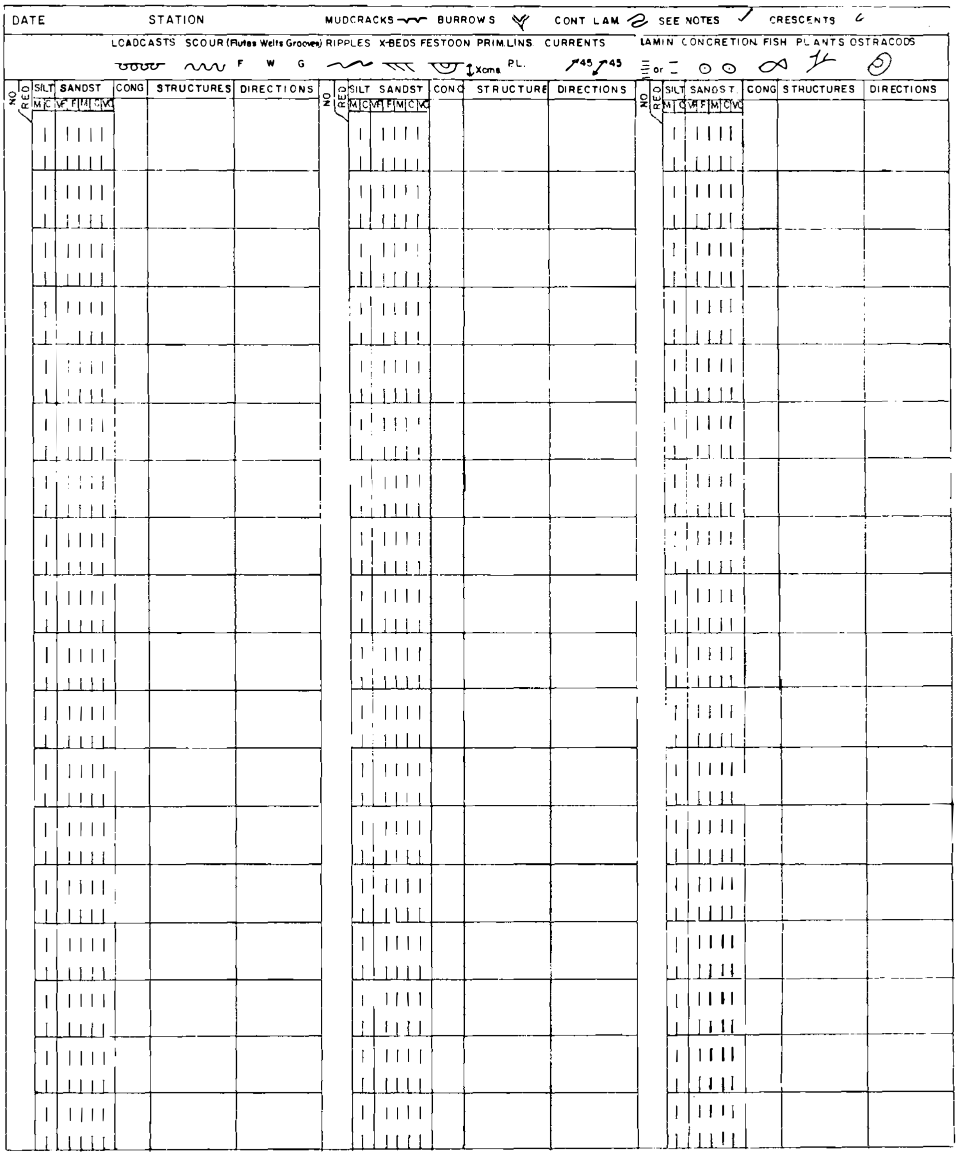

Fig. 3. *Field recording form used in this study. The actual size is 20 cm by 16.7 cm.*

Research Institute. For each locality the input consisted of the values of 1) the two geographical grid coordinates and 2) the variable (e.g. average grain-size). These data can be thought of as a three-dimensional scatter of points, one for each locality measurement, where the two horizontal dimensions represent the grid coordinates, and the vertical dimension represents the variable. The trend-surface part of the program specifies a planar or gently curved surface which most closely fits this scatter of points. We used a quadratic equation for this surface in all cases. COLE's iterative-fit technique then modifies this surface locally round each individual data point, to achieve a better local fit. The program repeated this iteration three times in all cases, each time improving the local fit. Detailed descriptions of this method are given in the publications cited above.

Compared with an overall trend technique, the main advantages of the fitting method from our point of view, are that:

a) the resulting surface fits the points better than an overall trend-surface. In some ways it approximates to an objective, repeatable, contour map.

b) the surface more satisfactorily deals with an elongated and irregular scatter of points. Local detail can often be easily distinguished from an overall trend.

When we come to interpret the features shown by the iterative-fit trend-surfaces, we can think of various possible explanations:

a) area-wide trends of lateral sedimentary variation.

b) more local trends of lateral sedimentary variation.

c) variation introduced by poor stratigraphical control. We shall describe below our three-fold subdivision of the Wood Bay Formation, which is based on fossils. These subdivisions vary from 400 to 1500 m in thickness; our study is based on the comparison of sections usually about 100 m thick and haphazardly located within the subdivisions.

The discovery that the area-wide trends shown by our surfaces are consistent from subdivision to subdivision, and from variable to variable, gives us confidence that area-wide trends are not being destroyed by vertical variation. However we cannot be so confident when we come to consider more local variations. In these cases we must base our interpretation on all the different evidence available. It will clearly not be worthwhile to examine even more local variation by considering the differences (or "residuals") between the surfaces and the actual values. Our data are simply not good enough for this.

STRATIGRAPHICAL FRAMEWORK

The "Wood Bay Series" was defined by HOLTEDAHL (1914) as the main red-bed unit of the Spitsbergen Devonian succession.

FØYN and A. HEINTZ (1943) proposed three subdivisions, mainly on vertebrate faunal grounds but using some local lithological distinctions.

FRIEND (1961) found these lithological differences difficult to apply across the area, and proposed a scheme of local rock units. Later, on the basis of more complete knowledge, FRIEND, N. HEINTZ and MOODY-STUART (1966) amalgamated these into one Wood Bay Formation, specifying some local members but no overall lithological subdivisions (Fig. 4). They proposed that the FØYN and HEINTZ subdivisions be retained as "faunal divisions".

We have used one vertebrate faunal boundary (FRIEND, HEINTZ and MOODY-STUART 1966) for widespread subdivision of the Wood Bay Formation (Fig. 5). This is the boundary between

1) *Kapp Kjeldsen faunal division:* species of *Gigantaspis* N. HEINTZ 1962, small species of *Doryaspis* WHITE 1935, (see N. HEINTZ 1968 and DENISON 1970 p. 37), near the top *Doryaspis nathorsti* (LANKESTER 1884), and

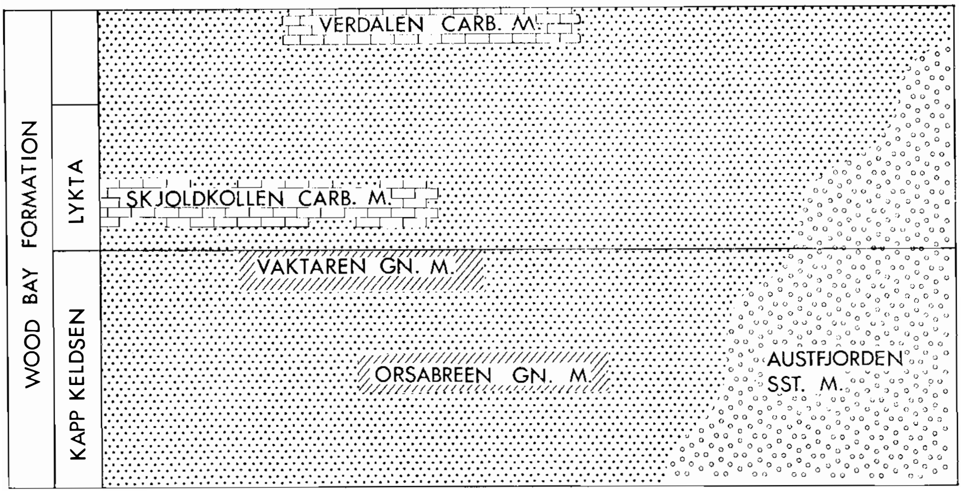

Fig. 4. *Faunal divisions (Kapp Kjeldsen, Lykta) of the Wood Bay Formation and lithological subdivisions. Carb. – Carbonate, Gn. – Green, M. – Member.*

2) *Lykta faunal division: Doryaspis nathorsti,* no *Gigantaspis,* no small species of *Doryaspis.*

For the purposes of this paper, we distinguish three informal units within the Wood Bay Formation, based on the recognition of this boundary (Fig. 6).

Lower: below the Middle, (up to 1500 m thick).

Middle: 200 m below to 200 m above, the boundary, (400 m thick).

Upper: above the Middle, (up to 1000 m thick).

This is the nearest we have been able to approach to an area-wide time subdivision.

The Wood Bay Formation ranges in age from basal Siegenian to basal Eifelian, (FRIEND 1961; ØRVIG 1969a), extending through the middle and upper parts of the Lower Devonian. Its deposition lasted, therefore, something like 20 million years (FRIEND and HOUSE 1964).

Symbols used in this paper

a = uplift block in isostatic model
b = down-dropped block in isostatic model
D = flow depth
Di = isostatic component of down-drop
E = erosion
f = friction factor
g = gravitational constant
Ha = height of surface of uplift block
Hb = height of downdrop block
p = fluid density
S = flow slope
Se = sedimentation
U = mean flow velocity
U_I = isostatic component of uplift
U_T = tectonic uplift
ω = power of stream flow
τ = bed shear stress per unit area

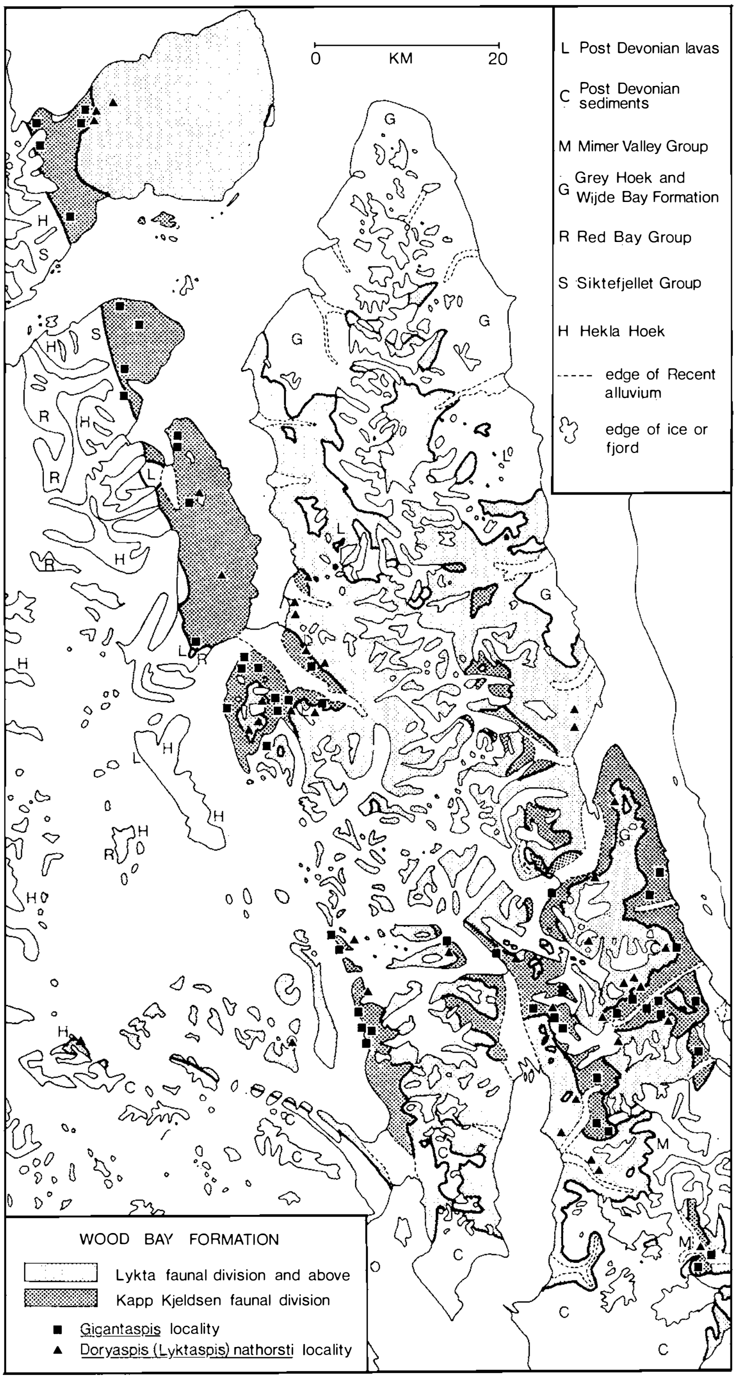

Fig. 5. *Map of the outcrop distribution of the Wood Bay Formation in northern Spitsbergen. Localities where critical fossils have been found by our parties are marked. The lowest (Kapp Kjeldsen) faunal division is distinguished from the higher ones. This is the standard map area, outlined on Fig. 1.*

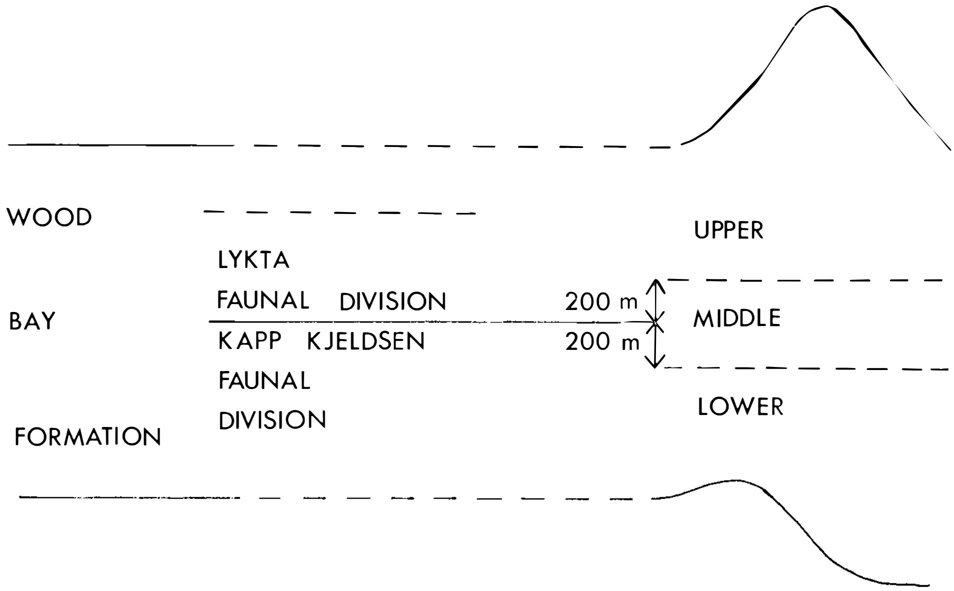

Fig. 6. *Informal time-rock units ("Lower", "Middle", "Upper") used in this analysis of the Wood Bay Formation.*

Palaeocurrent pattern

We have found the analysis of palaeocurrents (Fig. 7) to be our most powerful method.

The following structures, previously described by FRIEND (1965), were used:

Direction unambiguous: asymmetrical ripples
large-scale cross-stratification (foreset dips, trough orientation)
scoured sole structures (crescents, flutes)

Direction ambiguous: parting lineation
scoured sole structures (grooves, long welts)

Azimuths (projections of the directions on a horizontal plane) were measured in all cases, except where the tectonic dip was more than about 50°, when the pitch was measured. An original direction was estimated stereographically, making the reasonable assumption that the tectonic folding occurred about essentially horizontal axes.

For each locality, we have used all the measurements which we made in the field, without any selection or weighting. We are aware that our method of grouping a highly varied sample of flow vectors, is a crude one. ALLEN (1966 p. 184) has pointed to the systematic variation of flow vectors to be expected comparing one sort of structure with another. However, all our structures belong to his third and fourth orders of flow vector fields. For the purposes of this regional paper, we feel justified in grouping our structures together, and treating the vector mean

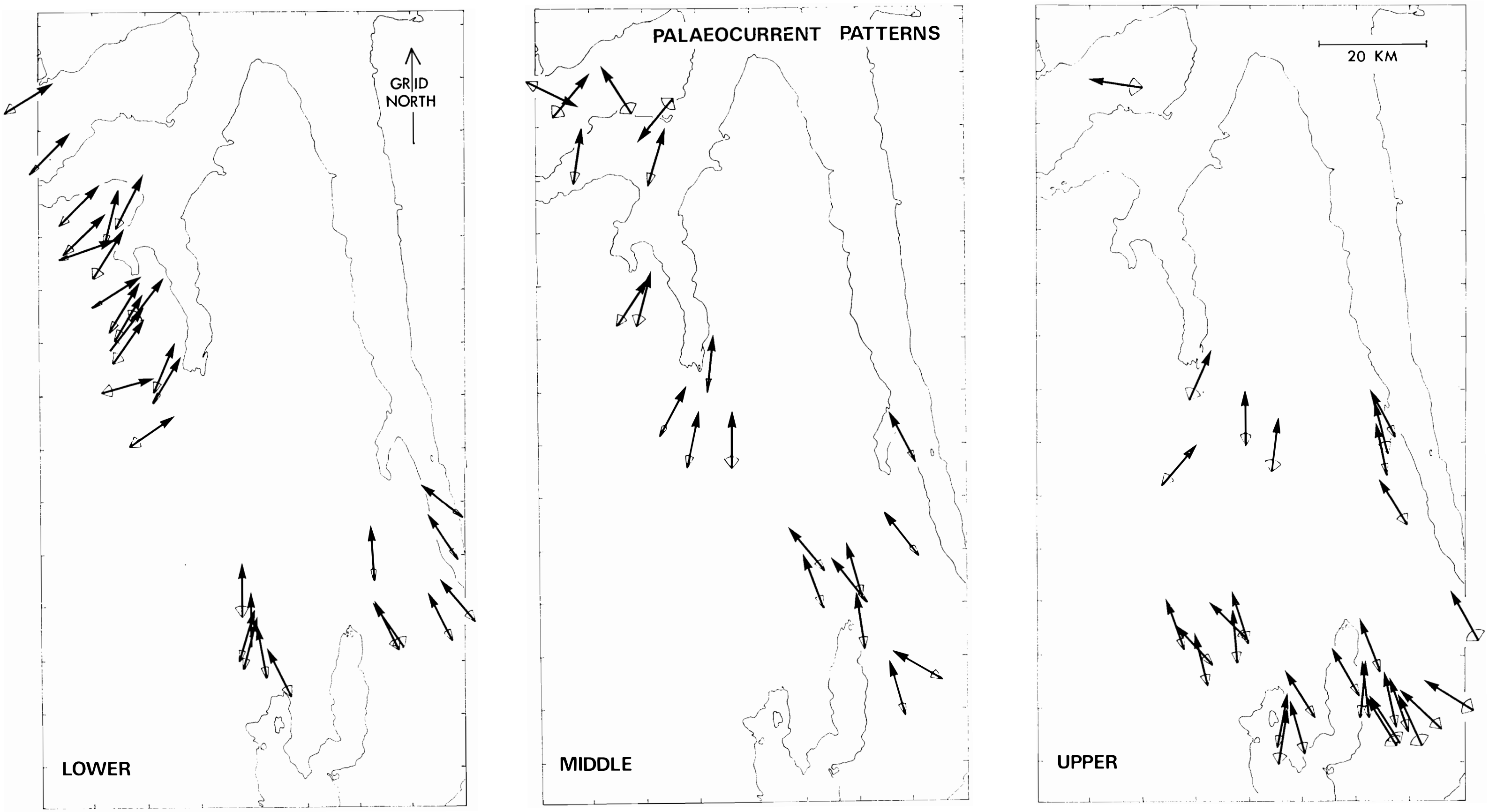

Fig. 7. *Standard map area (outlined on Fig. 1) for the three informal time-rock units, showing palaeocurrent patterns. At each locality marked by the point of the arrow, the vector mean is shown. The 95% confidence limits for the vector mean are plotted on the tail of each arrow.*

as a measure of the local palaeoslope. This has received recent support from studies of sedimentary structures in present-day streams. Both in meandering (BLUCK 1971) and braided (WILLIAMS and RUST 1969) streams, vector means of all the structures are good indicators of local palaeoslope.

A mean, and a measure of scatter, have been worked out for each locality, using the method devised by CURRAY (1956). In the case of ambiguous structures, the direction was chosen which was nearest to the direction of the most closely associated unambiguous structure.

Sandstone composition

Our routine examination of thin sections of sandstone samples from the Wood Bay Formation, showed us that the proportion of detrital grains of feldspar and rock-fragment (clay- or mica-rich) varied strongly from sample to sample. Using PETTIJOHN's classification (1957 p. 291), the samples would be protoquartzites, subgreywackes, subarkoses and arkoses. Variation in the proportion of detrital orthoclase present (from 0% to 38% of the quartz and total feldspar) is the most obvious regional variable. We therefore made a special study of it.

We stained the orthoclase feldspar in over two hundred thin sections, using the sodium cobaltinitrate method (HAYES and KLUGMAN 1959). We placed the samples in HF fumes for about 20 seconds, and then stained them for about 20 seconds. We identified and counted the minerals present at a minimum of six hundred points in each section.

Fig. 8 presents the orthoclase proportion expressed as a percentage (Appendix 1) of the total feldspar and quartz count. Comparison with the patterns of grain-size variation presented below, showed us that the areas richest in orthoclase are also the areas with the coarsest locality mean grain-sizes. We now set out to investigate this regional correlation.

The critical question to be answered was whether the feldspar content does actually vary with the grain-size of the sample. We had in mind the possibility that feldspar identification or diagenetic destruction might depend directly on the sample grain-size. To test this we roughly estimated the grain-size of each thin section by measuring the long diameter of ten successive grains in a microscope traverse. The mean of these ten, expressed in graticule divisions, was used as a grain-size estimate (Appendix 1). Grain-size and orthoclase content were then plotted for each sample (Fig. 9). Unfortunately the feldspar content is not normally distributed, many samples having no feldspar, and it is therefore not possible to assess quantitatively the degree of correlation by the usual statistical methods. However some correlation is visually apparent in Fig. 9. A statistical summary of this apparent correlation pattern is provided by a least-squares regression line (HOEL 1960 p. 145).

If the dependence of orthoclase content on sample grain-size explained all the variation in it, there would be no regional pattern left after we had applied a

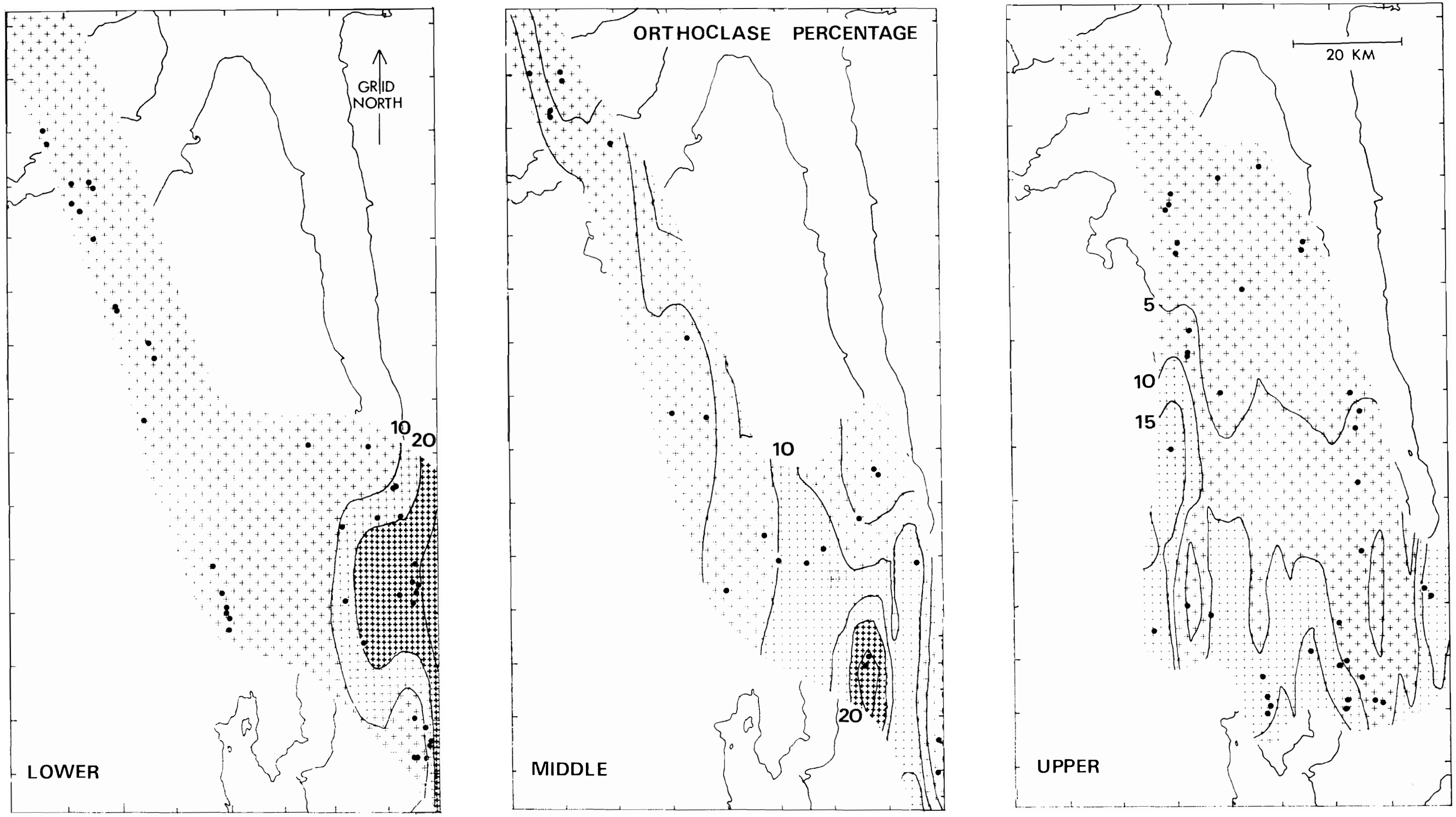

Fig. 8. *Standard map area (outlined on Fig. 1) for the three informal time-rock units, showing percentage of orthoclase in sandstone sample collected at each locality, marked by spot. Contours on iterative-fit quadratic trend surfaces have been plotted.*

grain-size correction. We tested this by expressing the orthoclase content of each sample as a percentage of the regression line at a particular grain-size. These percentages vary between 0% and 530%, and if plotted regionally (Fig. 10), patterns emerge which are broadly similar to those of the uncorrected values. We therefore conclude that feldspar content is not purely dependent on grain-size. Regional variation, independent of sample grain-size is also present. We presume that this reflects differences in the composition of the sediment supplied to the rivers. The general grain-size correlation then shows that the rivers supplied with most feldspar detritus also deposited higher proportions of coarse material.

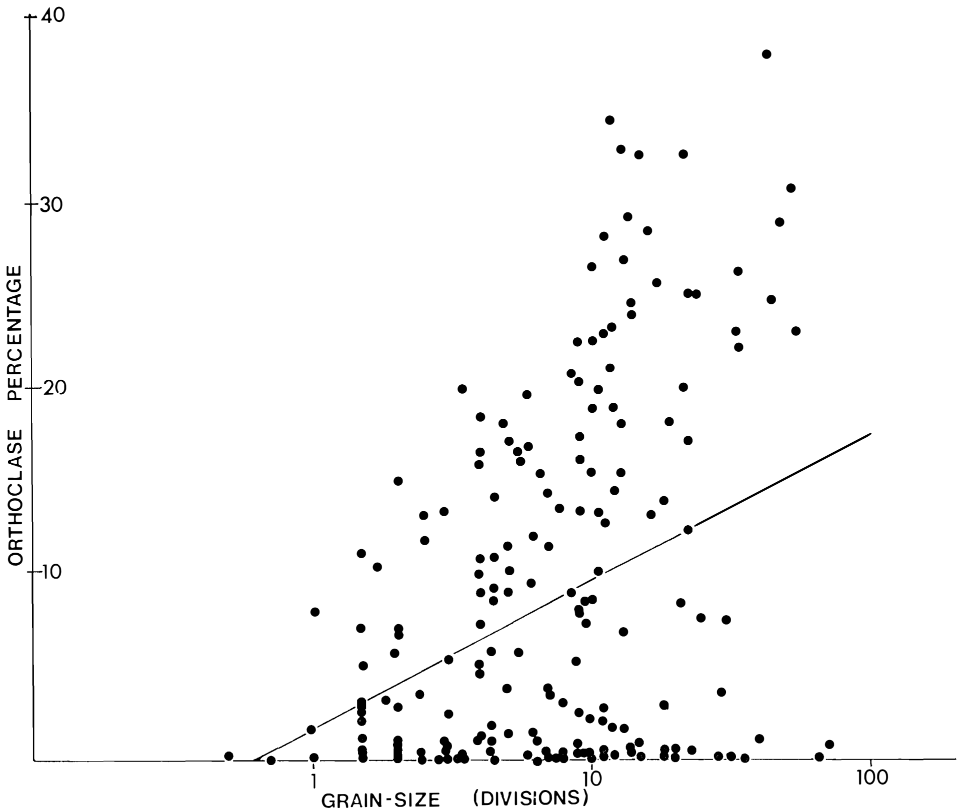

Fig. 9. *Plot of orthoclase content and grain-size for over two hundred sandstone samples. The orthoclase content is plotted as a percentage of the total feldspar and quartz point-count in thin section. The grain-size is measured, as described in the text, in graticule divisions (1 division = .015 mm). A least-squares regression-line is fitted to the scatter. This is the line of percentages on the logarithm of divisions.*

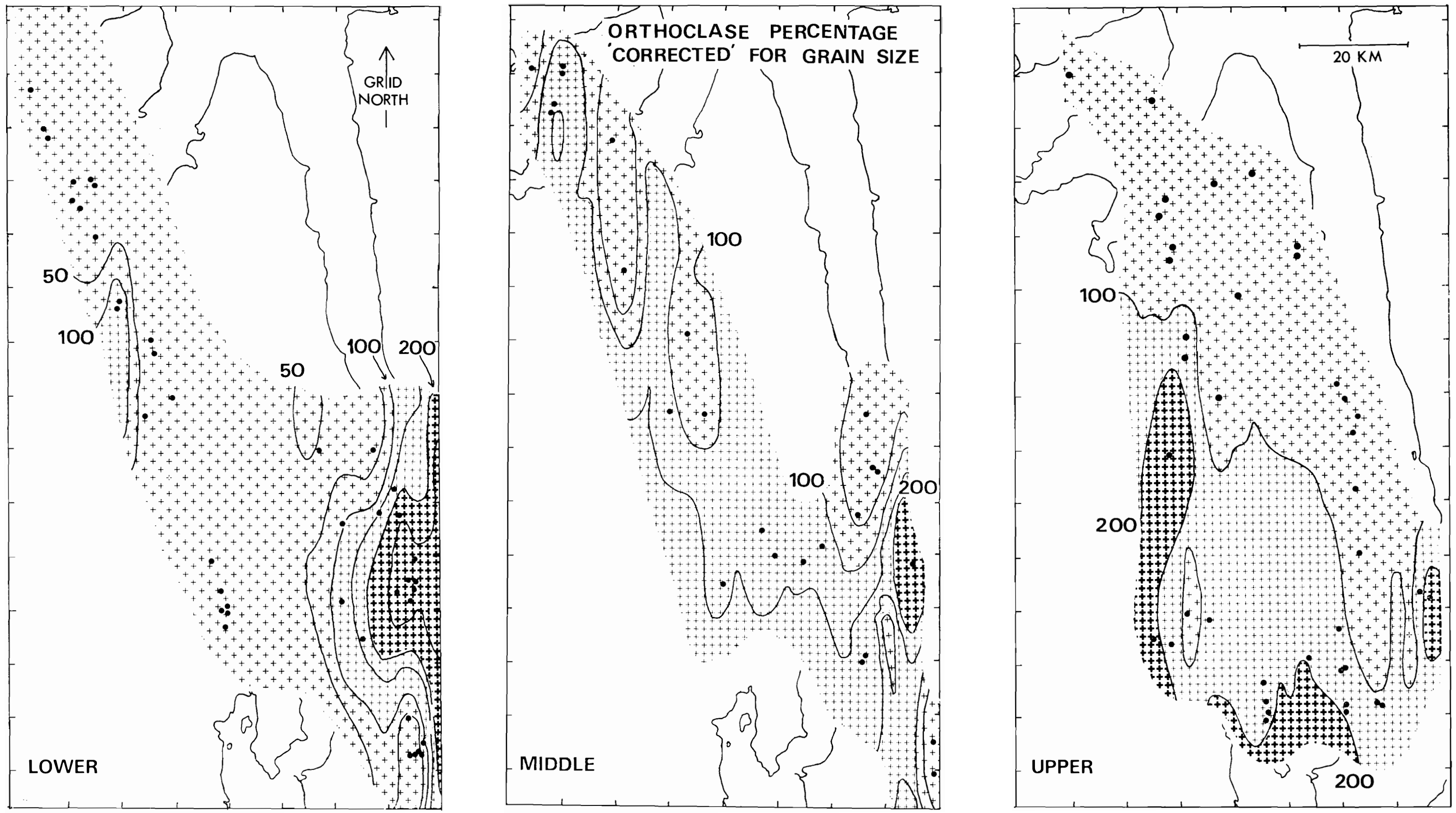

Fig. 10. *Standard map area (outlined on Fig. 1), for the three informal time-rock units, showing ortho-clase content 'corrected' for grain-size of the sample. The 'corrected' contents are percentages of the regression line value (Fig. 9) for the grain-size of the sample in question. Contours of iterative-fit quadratic trend surfaces are plotted.*

Grain-size variation

LOCALITY GRAIN-SIZE

As a first stage in considering variation of grain-size in the Wood Bay Formation, we have calculated a mean grain-size for each sedimentological section. Our method has been to note the grain-size at quarter metre intervals up the section, and determine the average of these grain-sizes. Plots of these data are presented in Fig. 11.

The most striking feature of these plots of average grain-size is the way in which eastern areas are consistently coarser than the central and western areas.

SEMI-CYCLES

Definition

Analysis in terms of fining-upwards cycles (or cyclothems) has become a valuable technique in the investigation of those fluvial formations that consist of successions of alternating siltstone and sandstone units. Examples of this technique have been common in recent papers on the Old Red Sandstone of Southern Britain (ALLEN 1964, 1965a, 1970), Eastern U.S.A. (ALLEN and FRIEND 1968; ALLEN 1970) and Spitsbergen (FRIEND 1961, 1965; MOODY-STUART 1966).

It was our object to analyse these cycles quantitatively on a regional scale. We therefore required a definition of the cycle which could be objectively applied. We have not attempted to define a cycle based on the simultaneous multivariate analysis of many characters (grain-size, structures, scale), although this would be an interesting study in itself. We have rather adopted a univariate approach by defining a cycle on the basis of one variable alone, grain-size. This approach is simple to apply and easy to interpret.

To distinguish these cycles from others, we have called them semi-cycles (ALEXANDER-MARRACK, FRIEND, and YEATS 1971). Our definition of a fining-up semi-cycle (F) is illustrated in Fig. 12. We apply the name "coarsening-up semi-cycle" (C) where a coarser set at the top of a fining-up semi-cycle is succeeded by an even coarser set. This is also illustrated in Fig. 12, where it can be seen that a coarsening-up semi-cycle may consist of one set only.

Maximum grain-size and sandstone percentage

The next feature we examined was the maximum grain-size in each semi-cycle. For this purpose we ignored the presence of pebbles, because, in the Wood Bay Formation, these rarely form more than thin and scattered zones in sandstones.

Fig. 13 shows the variation in maximum grain-size. It is apparent that the western areas generally contain the same maximum grain-sizes as the eastern areas, even though their average grain-sizes (Fig. 11) are consistently less.

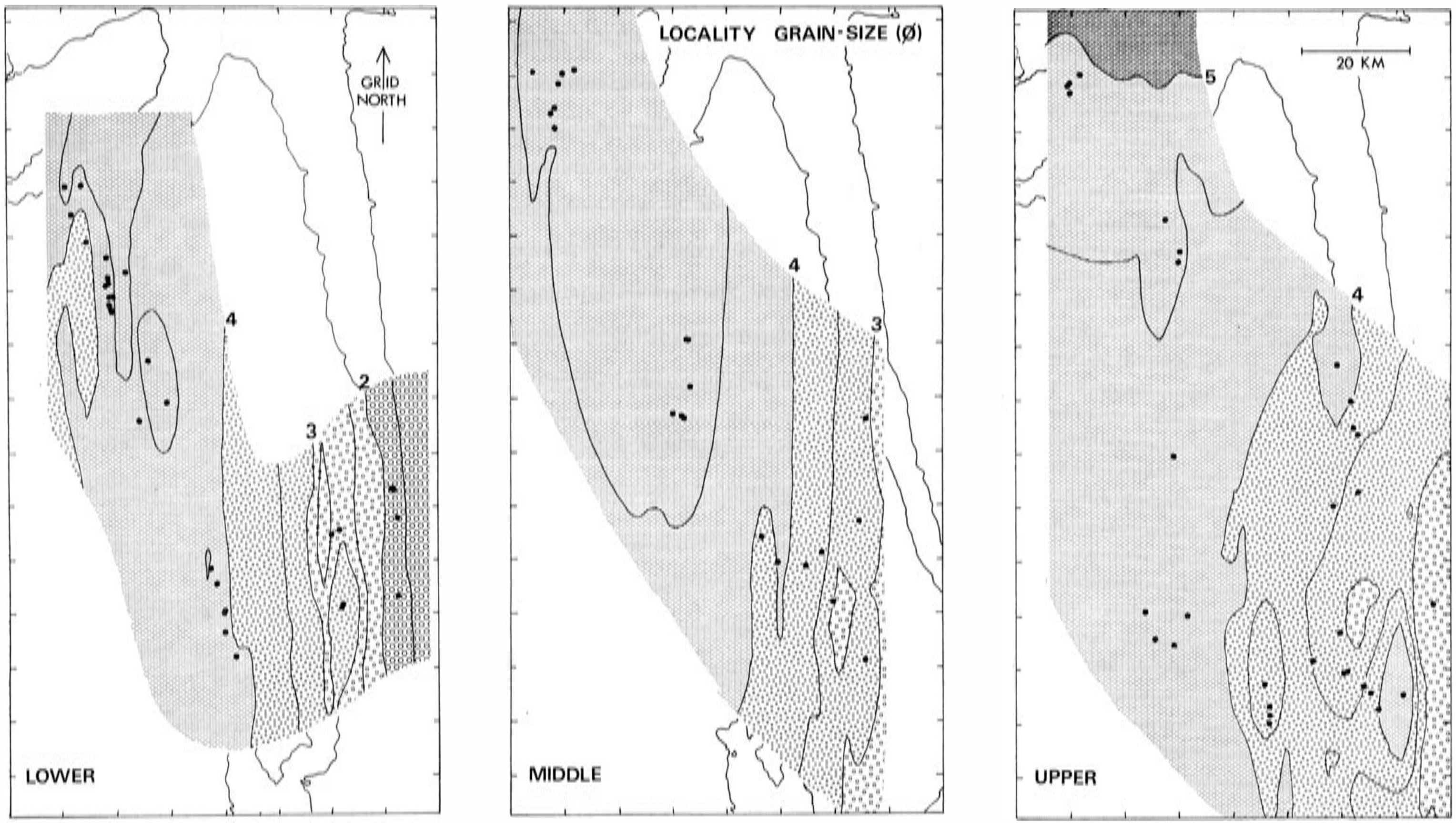

Fig. 11. *Standard map area (outlined on Fig. 1), for the three informal time-rock units, showing locality grain-size in Ø units. (> 4Ø is siltstone, 4Ø–3Ø very fine sandstone, 3Ø–2Ø fine sandstone, <2Ø medium and coarser sandstone). At each locality, marked by a spot, the grain-size calculated is the mean of the grain-sizes of each* 0.25 *m interval in the 'vertical' sedimentological section. Contours on iterative-fit quadratic trend surfaces have been plotted.*

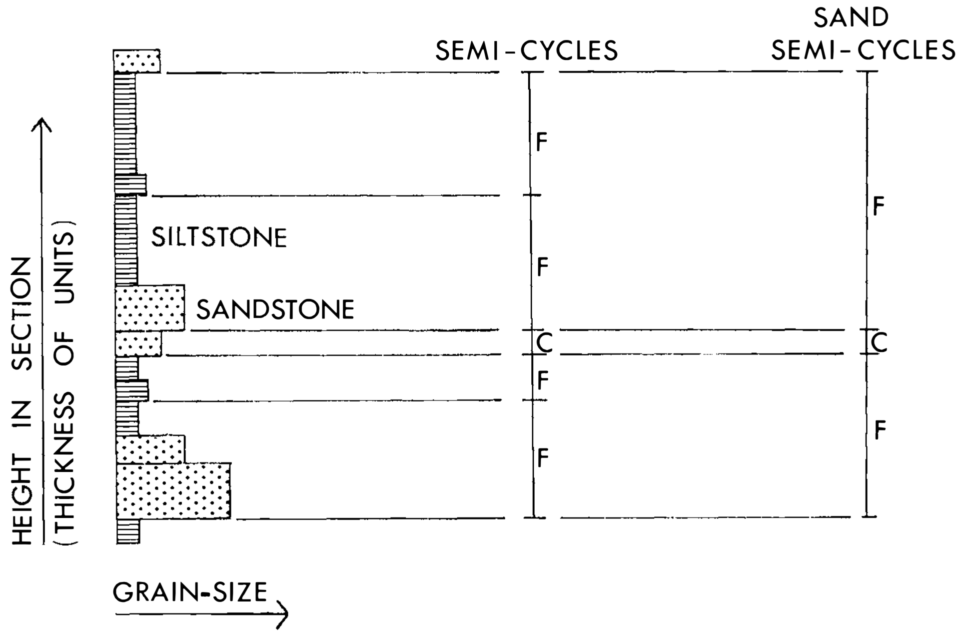

Fig. 12. *Diagram illustrating the definition of semi-cycles, and sand semi-cycles. F indicates a fining-up semi-cycle, C indicates a coarsening-up semi-cycle (in these cases, single sets).*

The explanation of this becomes clear when a plot of the percentages of sandstone in each semi-cycle is prepared (Fig. 14). The eastern areas (Austfjorden Sandstone Member) have a higher proportion of sandstone, and less siltstone than the western areas.

SAND SEMI-CYCLES

Definition

Fining-upwards cycles have generally been interpreted in terms of coarse member accumulation in a channel environment, and fine member accumulation in an overbank environment (e.g. Allen 1965b). The distinction between deposits of these two environments may, in some cases, be difficult to apply with certainty, especially where one of the two environments is not well developed, or where transitions occur (e.g. levée deposits, crevasse splay deposits). Because of this, we have found it helpful to make a distinction between bed-load deposits and suspended-load deposits. In the field, bed-load deposits are positively identified by their bed-form structures (asymmetrical ripples, cross-stratification, parting lineation).

Friend (1965 p. 46) published an analysis of the sedimentary structures in sets of different grain-size in some Wood Bay sections. This analysis suggests that a similar genetic distinction can be made, as a working approximation for regional analysis, between sandstone and siltstone sets. 80% of sets of medium siltstone grade were unlaminated or flat-laminated, and may, therefore, have been deposited from suspension. In contrast, 90% of sets of very-fine sandstone contain

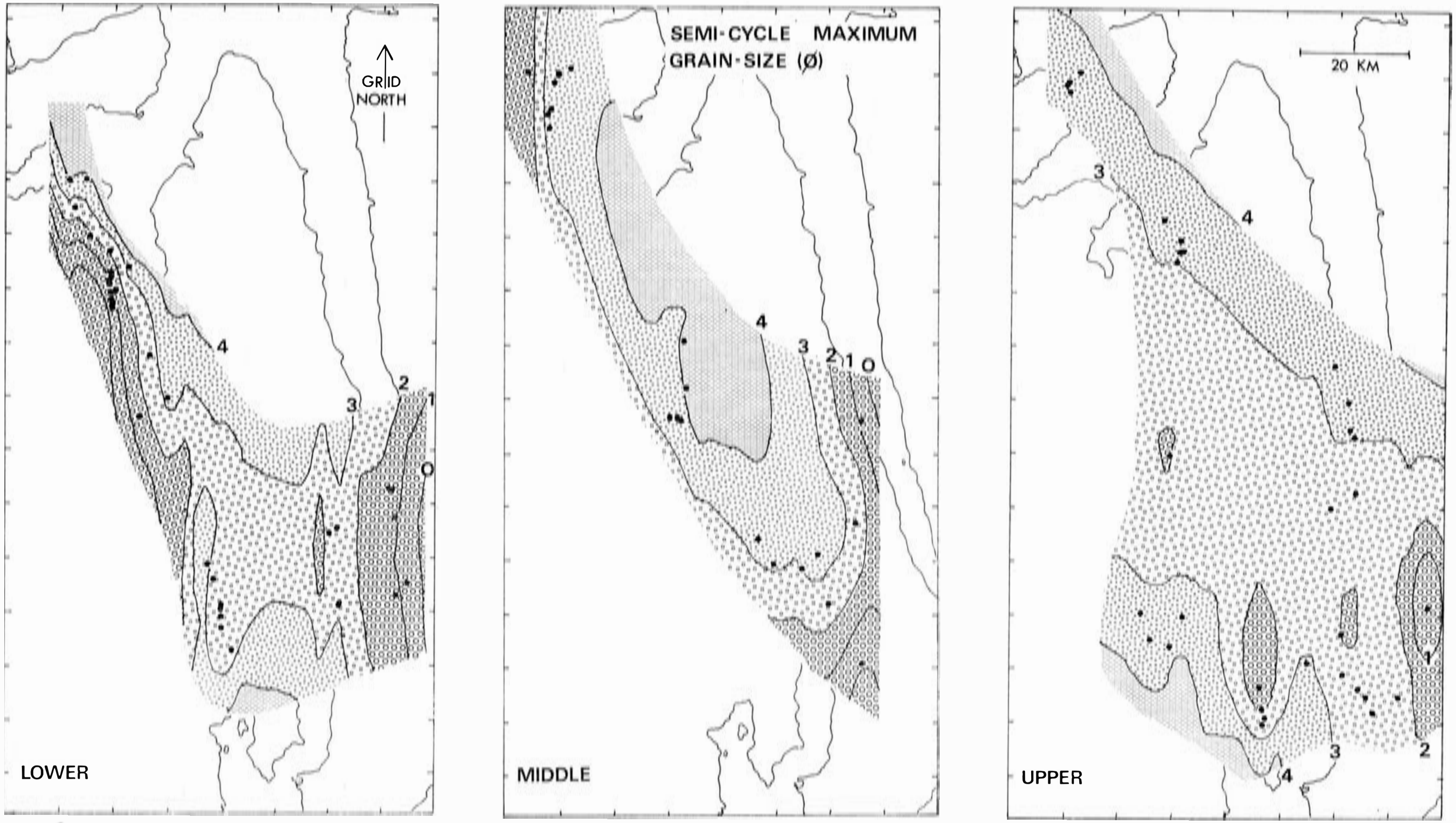

Fig. 13. *Standard map area (outlined on Fig. 1), for the three informal time-rock units, showing mean semi-cycle maximum grain-size. Each locality is marked by a spot. Ø units are used (described in caption to Fig. 11). Contours on iterative-fit quadratic trend surfaces have been plotted.*

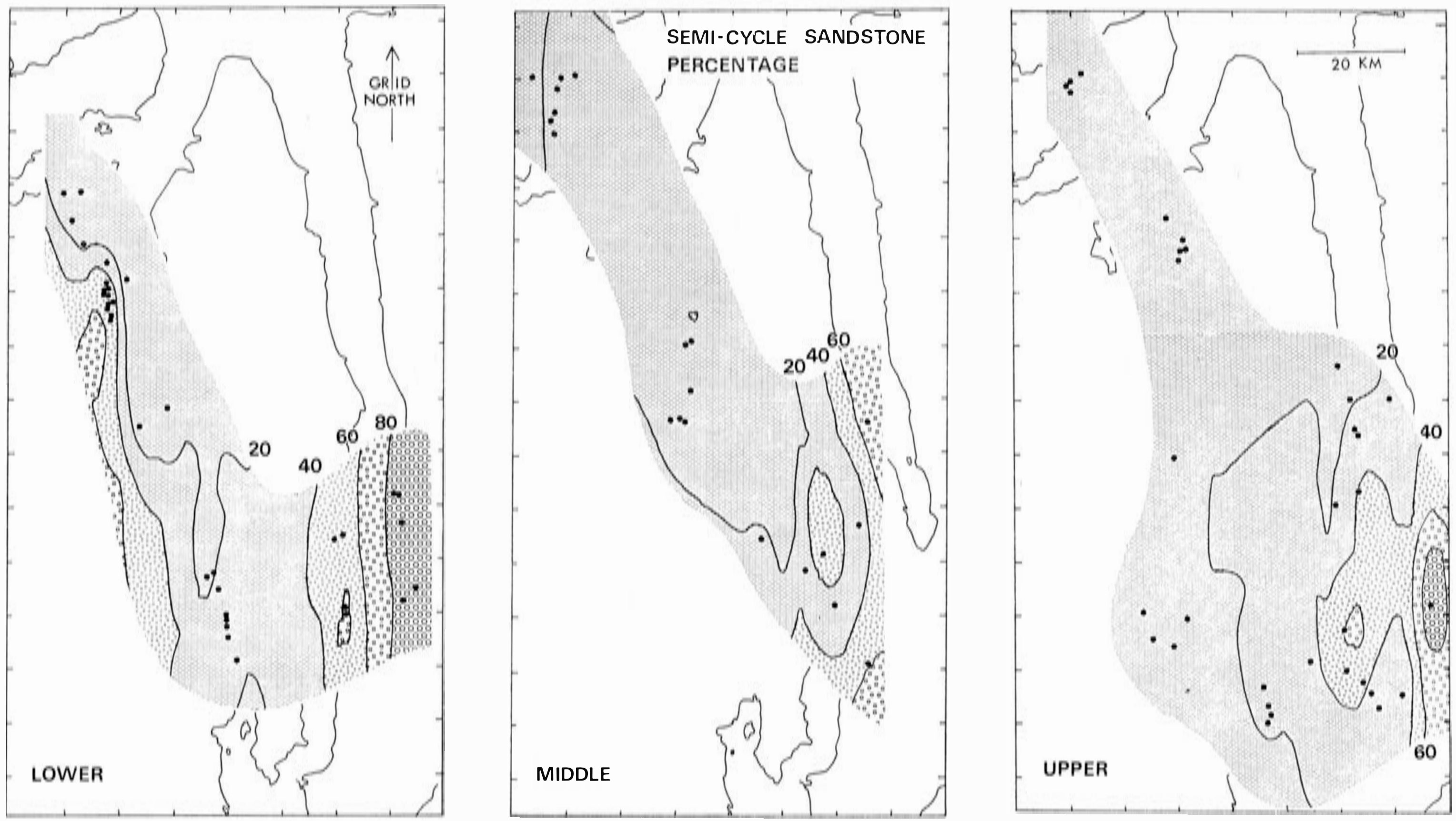

Fig. 14. *Standard map area (outlined on Fig. 1), for the three informal time-rock units, showing mean semi-cycle sandstone percentage. Each locality is marked by a spot. Contours on iterative-fit quadratic trend surfaces have been plotted.*

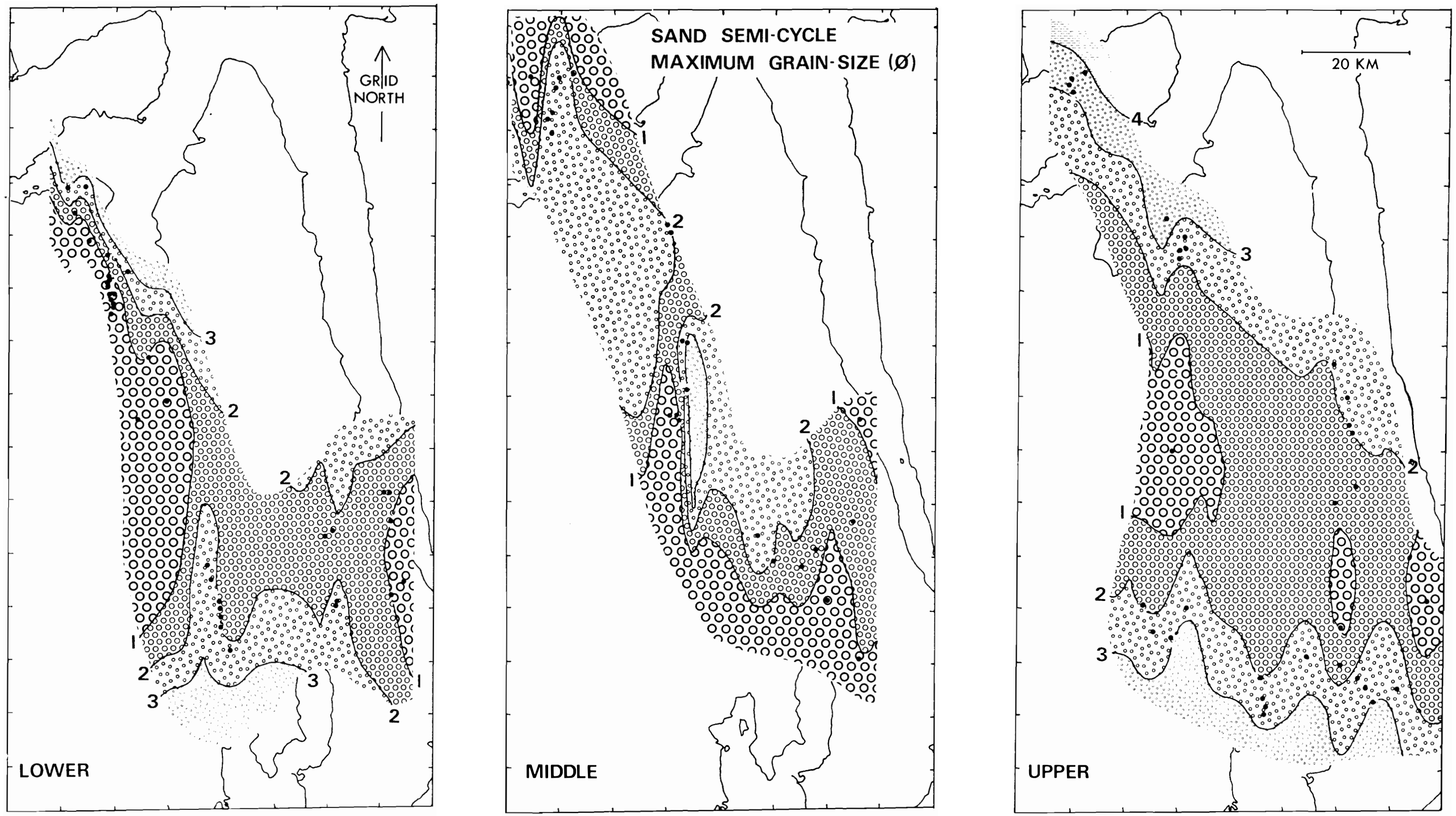

Fig. 15. *Standard map area (outlined on Fig. 1), for the three informal time-rock units showing mean sand semi-cycle maximum grain-size. Each locality is marked by a spot. Ø units are used (described in caption to Fig. 11). Contours on iterative-fit quadratic trend surfaces have been plotted.*

small and large cross-stratification or parting lineation, and were, therefore, deposited from bed-load.

To use this distinction, we have defined another kind of semi-cycle, that which contains one or more sets of sand-grade (Fig. 12). These "sand semi-cycles" may either coincide with the semi-cycles considered above, or they may include two, or more of them. They have the special feature that it is reasonably certain that they included an episode of bed-load deposition.

Maximum grain-size

The pattern of variation of maximum grain-size of the sand semi-cycles is plotted on Fig. 15. The patterns for "Lower" and "Upper" times are very similar to those for the general semi-cycles (Fig. 13). However the pattern for "Middle" times is more complex, and particularly different in northern areas. This is because of the rareness of sand members in this region, and the variability of grain-size of the few that do occur.

Thickness

The pattern of thickness of the sand semi-cycles is plotted on Fig. 16. It shows a range of averages from 4 m (or less) to 16 m (or more). The sand semi-cycles become thicker in a generally downstream (Fig. 7) direction.

Thickness of sand member

The variation in thickness of the sand member is plotted on Fig. 17. In "Upper" rocks this variation is minor, and no overall pattern can be discerned. In "Lower" and "Middle" rocks, there is consistent thinning of sand members downstream (Fig. 7), from a thickness of 4 or 6 m (similar to that of the whole semi-cycle) to thicknesses considerably less than 1 m.

Check on downstream trends of variation

To check on the variation which we have just deduced from the patterns of iterative-fit trend-surfaces, we have (Fig. 18) selected groups of localities along palaeocurrent streamlines. We have then averaged for each group of localities, the characteristics of all the sand semi-cycles. On Fig. 18 we have plotted the actual average values against the values of the trend surfaces. It will be seen that these correspond closely.

Variation of individual sand semi-cycles

Any study of a feature which depends solely on the averages of individual characteristics, runs the risk that the average may conceal important variation. We have therefore plotted for four of the groups of localities, the characteristics of individual semi-cycles (Fig. 19). Against semi-cycle maximum grain-size, we have plotted thickness of sandstone member, and thickness of siltstone member. The sum of the sandstone member and the siltstone member (if present), is the sand semi-cycle thickness which we analysed above.

This allows us to make a model of different sorts of sand semi-cycles, and their probable lateral variation (Fig. 20). But we would stress that in no case have we been able to follow an individual sand semi-cycle far enough to be able to see these variations.

In the "Lower" western streamline (LW on Figs. 18–20) the commonest cycles consist of less than 1 m of sand, either of very-fine or very-coarse (and pebbly) sand. These sand members sometimes rest directly on each other, but there is more often a siltstone member and these may be up to 10 m thick. A few kilometres downstream (LW2) the sand members are missing altogether, and the entire section is made of siltstone.

In the "Upper" eastern streamline (UE on Figs. 18–20), in the upstream area (UE1), cycles consist of up to 5 m of sandstone, varying from fine to very-coarse and pebbly in grade. No siltstone occurs. Further downstream (UE 2 and 3), the sandstone members tend to be thinner and siltstone members are increasingly common. More sand semi-cycles have finer sand grain-sizes, so we conclude that some of the sand members do become finer-grained downstream.

Alluvial environments

THE ALLUVIAL VARIABLES

Kennedy (1971) analyses the very complex array of variables which make up an alluvial system. He distinguishes between *independent variables,* whose values are imposed externally on the system, and *dependent variables,* whose values are controlled, in turn, by the values of the independent variables. In our Spitsbergen study, we want ultimately to assess the independent variables.

Whether a variable is independent or dependent will depend, to some extent, on the nature of the system and on the period of time over which it is active. For instance, in some flume experiments, the slope is determined by the experimenter, i. e. externally and independently. In some natural streams, over a short period of time (days to months), the slope is also determined independently, because it has been inherited from the conditions of the previous major flood. However, over a longer period of time (several years), the slope of all natural alluvial streams is a dependent variable controlled by the independent variables, water and sediment discharge (Kennedy 1971 p. 113). This long term adjustment of slope is the core of the idea of a graded river (Mackin 1948).

In our regional work, our interest is mainly with the long-term balance of the variables of river systems. We shall therefore be concerned with the following pattern of independence and dependence (Kennedy 1971 p. 113):

Independent variables:	water discharge
	sediment discharge
Dependent variables:	alluvial grain-size
	channel cross-section and plan
	friction factor
	strength of river flow
	slope

We shall first consider the dependent variables, in turn, below.

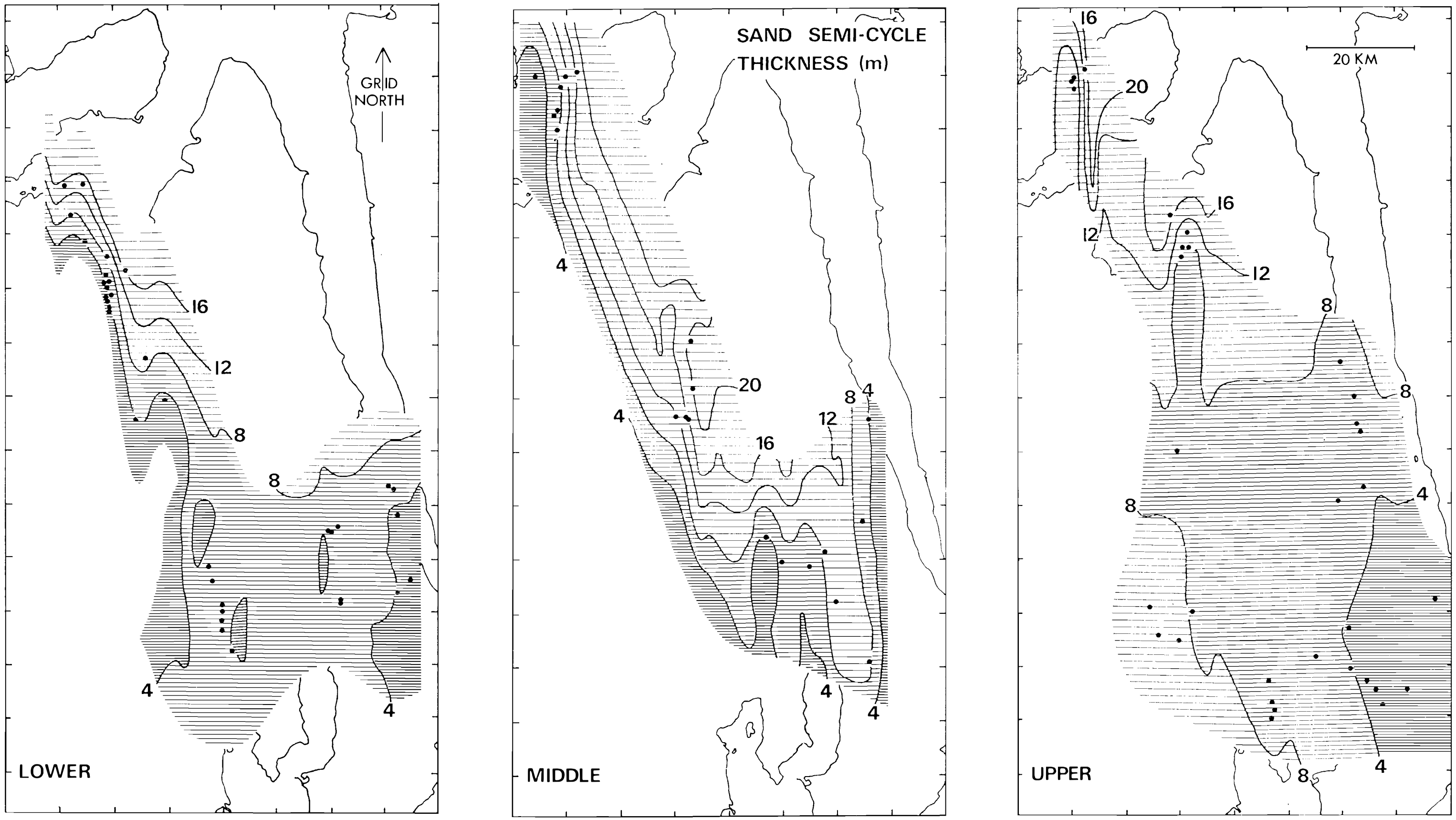

Fig. 16. *Standard map area (outlined on Fig. 1), for the three informal time-rock units, showing mean sand semi-cycle thickness in metres. Each locality is marked by a spot. Contours on iterative-fit quadratic trend surfaces have been plotted.*

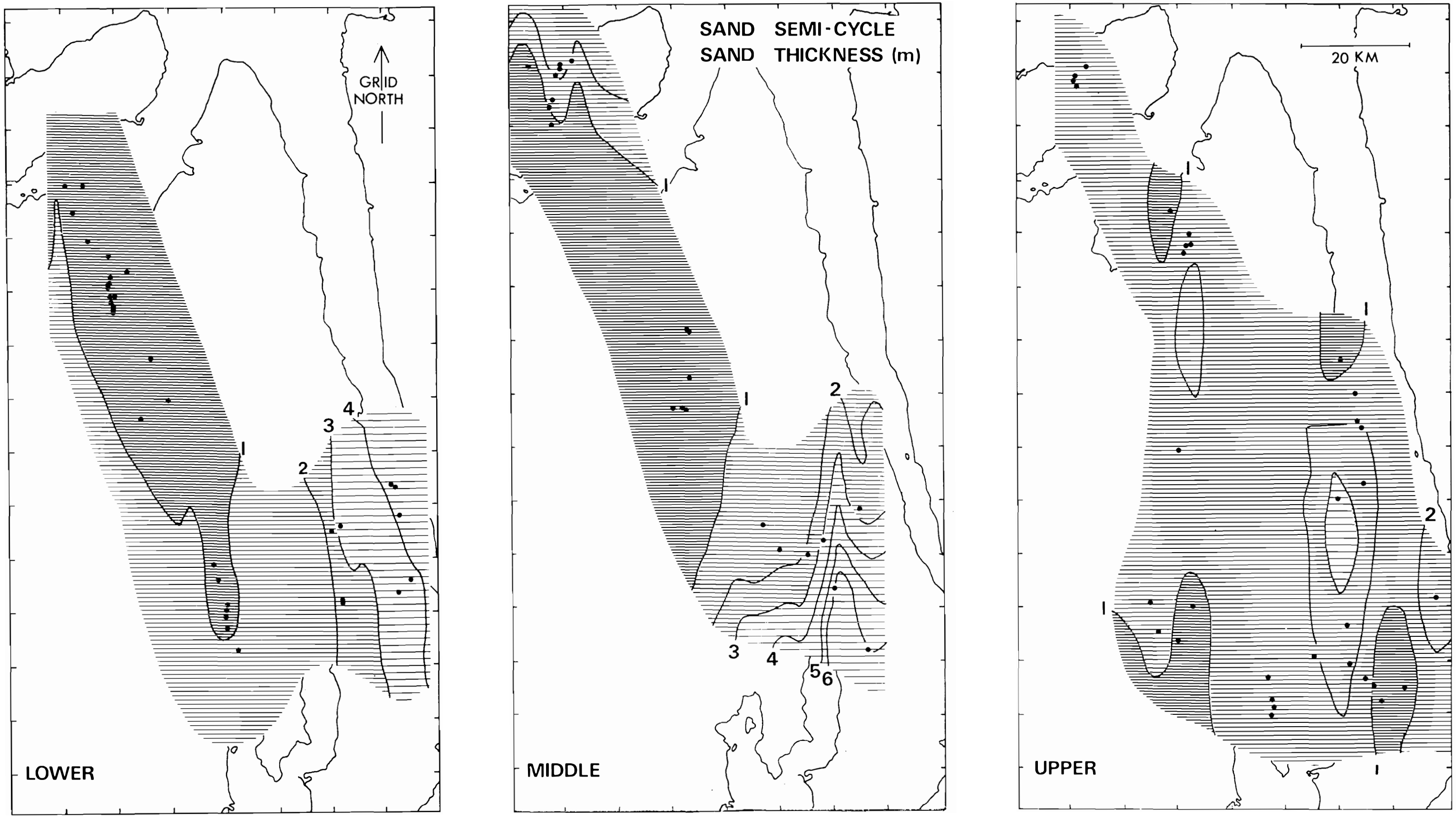

Fig. 17. *Standard map area (outlined on Fig. 1), for the three informal time-rock units, showing, in metres, the mean of the thickness of the sand units in each sand semi-cycle. Each locality is marked by a spot. Contours on iterative-fit quadratic trend surfaces have been plotted.*

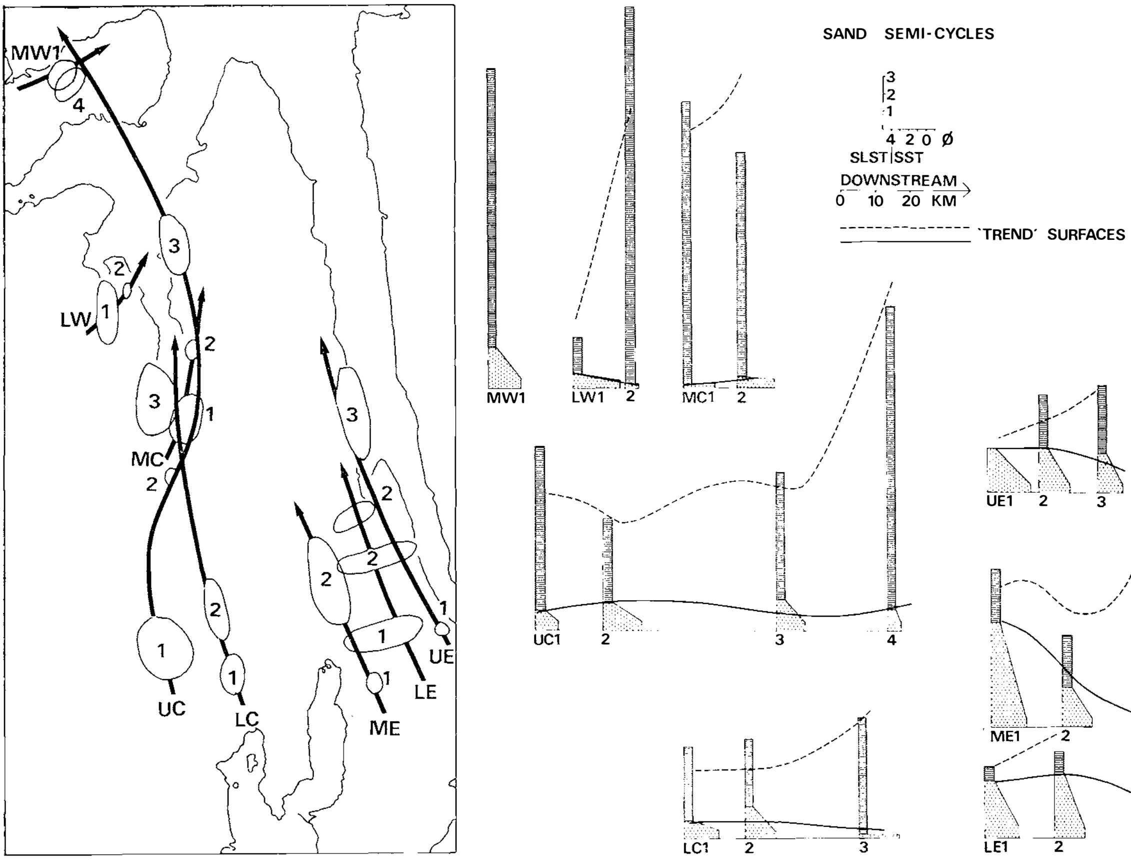

Fig. 18. *Study of downstream variation of sand semi-cycles. Data from localities is grouped from Lower (L), Middle (M) and Upper (U) informal time-rock units. Western (W), central (C) and eastern (E) streamlines are defined using the palaeocurrent trends of Fig. 7. From all the sand semi-cycles in each group, actual averages are plotted for total thickness, sand thickness and maximum grain-size. The corresponding trend surface values for the centre of each group are also plotted.*

ESTIMATING THE DEPENDENT VARIABLES

Alluvial grain-size

The grain-size of alluvium is partly controlled by processes acting in the source area. These processes determine the amount of sediment supplied and its maximum grain-size. In the alluvial area the grain-size is then further influenced by processes of local sorting, often involving many cycles of transport and deposition.

We have already distinguished between siltstones, mainly deposited from suspension, and sandstones, mainly deposited from bed-load. It is a remarkable feature of the bed-load deposits of the Wood Bay Formation that they have so limited a range of grain-size (very fine to medium sand grades, Fig. 15). Pebbles are rare and usually of locally derived, intraformational material. Conglomerates hardly exist. Our knowledge of modern rivers allows us to make a (general) estimate of stream strength based on this fact. This will be done below.

More subtle distinctions of grain-size, within the sand grade, may be due to a

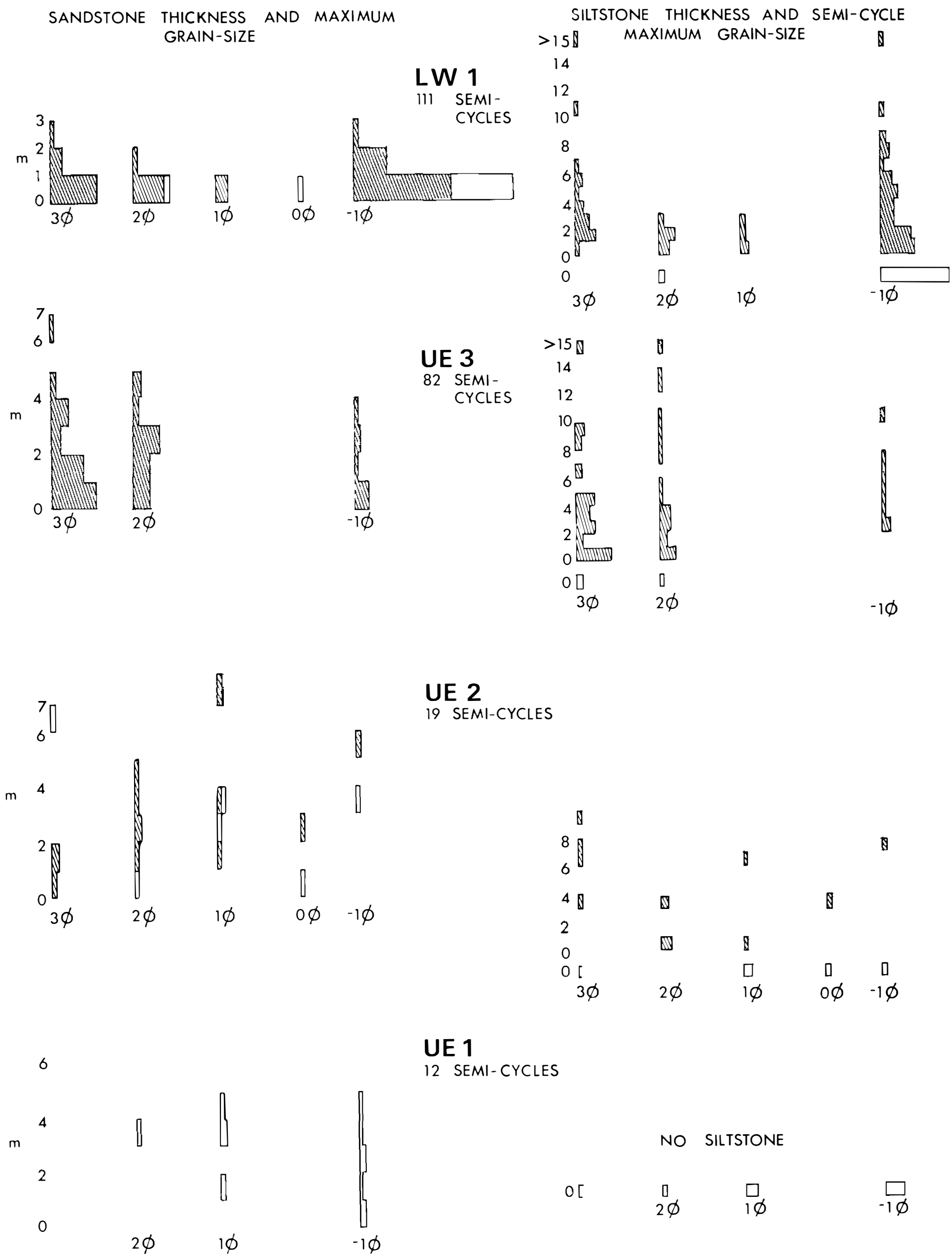

Fig. 19. *Diagrammatic plot of data for every sand semi-cycle measured in some of the groups of localities defined in Fig. 18. The data are presented as frequency histograms, showing the thickness and grain-size of all the sandstone and siltstone-members of each semi-cycle. Members of fining-up semi-cycles are shaded. Members of coarsening-up semi-cycles are clear.*

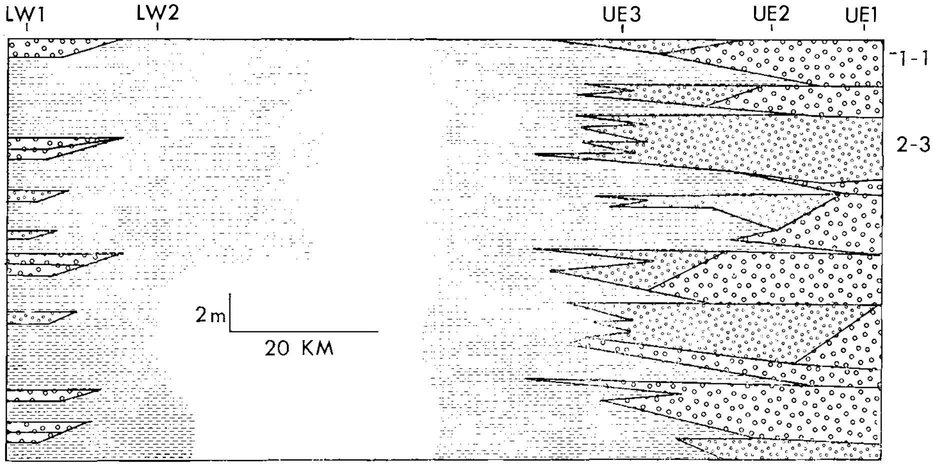

Fig. 20. *Hypothetical vertical-plane section showing lateral variation of sandstone units, and their relationships to siltstone units, in the western area groups (LW1 and LW2) and the eastern groups (UE1, UE2 and UE3) which are located in Fig. 18. These relations are characteristic of the eastern and western river systems.*

number of local processes of selective erosion, transport and deposition of the hydrodynamically different particle fractions.

Possible processes are:

a) sorting on river bar faces. Lateral components of river flow, weaker in shallow water, balance downslope components of particle weight (ALLEN 1970).

b) downstream sorting. Different size-fractions of bed-load travel at different rates, and are buried differentially in ripples (MELAND and NORRMAN 1969). Some fractions may temporarily travel as suspended load, during peak floods. They will therefore travel faster and further than coarser fractions which travel entirely as bed-load (BRUSH 1965).

Channel measurements

Channel type. — In his work on present-day rivers and their alluvium, SCHUMM (1968a, p. 1579) distinguished a *bed-load* type of channel. He defined channels of this type as those having less than 5% of silt or clay (suspended-load deposits) in a systematic sampling of their walls and floor (their perimeters). In terms of the rivers, rather than their perimeters, SCHUMM thought that they carry at least 11% of their total sediment load as bed-load. It is because of the difficulty of measuring bed-load, that SCHUMM's definition must, in practice, be based on perimeter sampling.

SCHUMM was able to generalise about channel cross-section and plan (1968a, p. 1580). Bed-load channels are characterised by great width compared with their depth (ratio about 60). They are almost straight (sinuosity 1.0 to 1.1) and are often braided, having extensive alluvial islands at low stage.

The deposits of a series of these bed-load channels accumulate as thick successions of sand or gravel (SCHUMM 1968, p. 1582) with only rare and thin beds of suspended load material. Although suspended-load sediments may accumulate as superficial veneers in years of low peak discharge, they tend to be destroyed in the wholesale reworking of the underlying cohesionless bed-load material which occurs during major floods.

On the basis of the grain-size of its channel perimeter, we would regard the Brahmaputra River (COLEMAN 1969) as of bed-load channel type. This is supported by its low sinuosity and braiding. Although there are extensive "floodbasins" adjacent to its braided channels, COLEMAN (1969, p. 233) stated that "Because of the large number of major and minor channels within the Bengal Basin and their active migrating nature, floodplain deposits form only thin veneers capping channel sands and silts".

McGOWEN and GARNER (1970) classify the Amite River (Louisiana) and the Colorado River (Texas) in SCHUMM's bed-load type. The banks of these rivers are unusually strongly stabilized by vegetation, so they are not braided, and have sinuosities of 1.4 to 1.75. The floodplain deposits are mainly of bed-load material (p. 89), like those of Bijou Creek (McKEE, CROSBY and BERRYHILL 1967).

It seems clear to us that the dominantly sandy deposits of the south-east (Austfjorden Sandstone Member) of the Wood Bay Formation were deposited by rivers of SCHUMM's bed-load type. This carries with it implications about channel cross-section and plan.

SCHUMM (1968a, p. 1579), also defined *mixed* and *suspended-load* types of channels. These have respectively between 5 and 20%, and more than 20%, of suspended-load sediments in their channel perimeters. He suggested that these channel perimeter features correspond to 3 to 11%, and less than 3% bed-load in the total sediment load.

As examples of the "valley fill" or floodplain stratigraphy of a mixed-load channel, SCHUMM (1968a, p. 1582) gave the "prior stream" channels of the Murrumbidgee River. Channel forms were cut at different times in a Pleistocene fluvial and aeolian formation, and were then plugged by the aggrading streams. Each plug consists of a lens of sand or gravel, decreasing in width upwards from its maximum, because the depositing channel decreased in size and deposited increasing amounts of cohesive suspended-load material. During its earlier and larger stages, each channel was of bed-load type (SCHUMM, 1968b, p. 36), but as time passed, it became increasingly a suspended-load channel, presumably passing through a mixed-load stage.

The present Murrumbidgee provides SCHUMM's example of the sort of alluvial succession built by a suspended-load type of channel. This is depositing an alternation of sand and clay members. "Lateral migration of this suspended-load channel leaves a sheet of channel sand which is overlain by point-bar deposits of lateral accretion and overbank deposits. Only about one-third or less of this deposit will be composed of sand or coarse sediments" (SCHUMM 1968a, p. 1582).

We do not feel confident that we can distinguish between the deposits of mixed and suspended-load types in the Wood Bay Formation. SCHUMM's examples allow us to feel sure that any section composed of sand and silt semi-cycles, was formed by mixed or suspended-load channel activity, compared with a sand only section which was formed by bed-load channel activity. Within the sand and silt sections around Woodfjorden, MOODY-STUART (1966) distinguished between high and low sinuosity channel deposits, and SCHUMM (1968a, p. 1582) suggested that they might correspond to suspended and mixed-load channel types. In our regional study we are not able to apply this distinction widely to our body of general data. We therefore group together *mixed*-or-*suspended-load* channel deposits, and characterise them by single values of the channel measurements. In contrast to the bed-load channels, they have a smaller width/depth ratio (10), and higher sinuosity (2) (SCHUMM 1968a, p. 1580).

Cycles. – We have explained above our decision to define a "semi-cycle" for analytical purposes, purely in terms of vertical variations of grain-size. We now come to consider possible origins for these semi-cycles.

BEERBOWER (1964) described *allocyclic* sequences in which the cyclicity results from disturbances outside the sedimentary system (alluvial part of system in our terminology) "by changes in discharge, load and slope". All these changes would produce characteristically widespread changes in the sedimentary basin.

(1) *Changes of water discharge*. The "prior streams" of the Murrumbidgee River produced fining-upwards semi-cycles (SCHUMM 1968b, p. 30). The fining-upwards was partly the result of decreasing peak discharge and reduction of sand load due to change of climate.

(2) *Change of sediment-load*. SCHUMM (1968a) explained how sediment may accumulate in source area valleys, to be flushed out intermittently by exceptional (say 1000 year) floods and deposited in an alluvial area.

(3) *Change of slope* (or base level). A rise in base level first decreases the slope nearest to that level. This results in a decrease in flow strength, and deposition of sediment. The Recent deposits of the Mississippi River are examples of a large single fining-upward semi-cycle due to this cause. In the case of the Mississippi, the change from erosion to gravel, and then sand, deposition has taken place over about 30,000 years (FISK 1952, p. 66). The river Cam, near Cambridge, has a similar sequence: erosion surface – sand and gravel – silty clay, which again reflects post-glacial sea level rise (SPARKS and WEST 1965, p. 35).

If the cycles can be shown to be local, not reflecting basin-wide changes, they are of BEERBOWER's *autocyclic* type. These are "generated purely within the sedimentary prism". Subsurface plots of the cyclic deposits of the alluvial Rechna Doab, West Pakistan, provide a clear example of local cycles which obviously result from movements of the rivers of the area (KAZMI 1964). We have no doubt that the cycles of the Wood Bay Formation are of this local type. Fig. 21 presents a rare example of sections which are known to be lateral equivalents (i.e. not separated by faulting). The cycles are highly impersistent.

We shall therefore consider interpretation of Wood Bay cycles in terms of local factors. In Fig. 22, we present a range of typical successions in the Wood Bay Formation. We have classified the semi-cycles in three categories (A, B, C below and in Fig. 22).

(A) *Single-set sandstone and siltstone*. The presence of a silt member, assuming it to be thick, or laterally persistent, is evidence for a significant episode of deposition from suspension in fluvial overbank, or lacustrine conditions. The fact that, in this category, the episode of bed-load deposition resulted in a single set of one

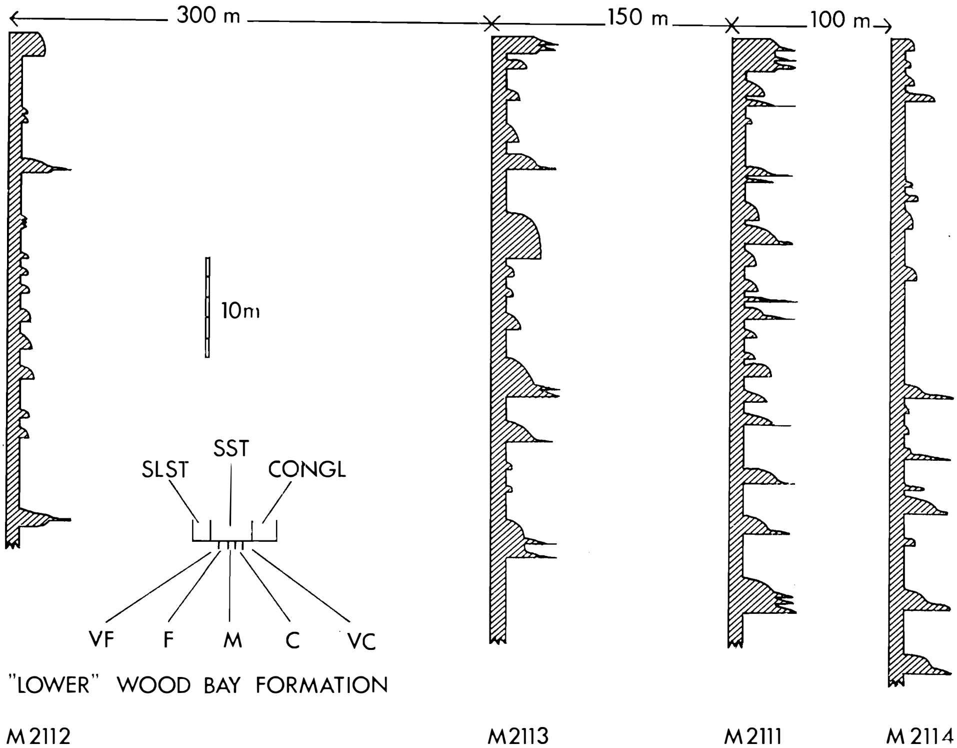

Fig. 21. *Four 'vertical' sedimentological sections (grain-size against position in log) whose relative positions are known. No significant faults separate these sections, so they provide some idea of the lateral variation in number and grain-size of sandstone members.*

grain-size and one type of internal structure, means that the episode was short-lived, probably the duration of a single flood. Single sets of this sort have been deposited by particular floods of the Connecticut River (JAHNS 1947). They are a feature of the lower parts of the fan deposits of the northern Sahara (WILLIAMS 1970). They are also a feature of many levée deposits, in which sand members accumulate during one or more floods as crevasse splays. In the deposits of the Brahmaputra (COLEMAN 1969 p. 230), the levée sand members are thickest close to the main channel, and thin progressively away from it.

(B) *Multiple-set sandstone and siltstone.* In this category the presence of a silt-member again points to a significant episode of suspended-load accumulation. However it differs in that the sand member consists of more than one sedimentation event. By definition the sets become finer in grain upwards, and are similar to the point bar deposits of some meandering streams (e.g. BERNARD and others 1970). This fining upwards has been explained by ALLEN (1970). It results from the balancing of the component of weight of the various particles down the bar surface, against an upsurface component of river flow which decreases as the water becomes shallower. The sand-member thickness corresponds directly to the bank-full depth of the channel.

In a cyclic sequence of this sort, the change from an episode of major suspended-

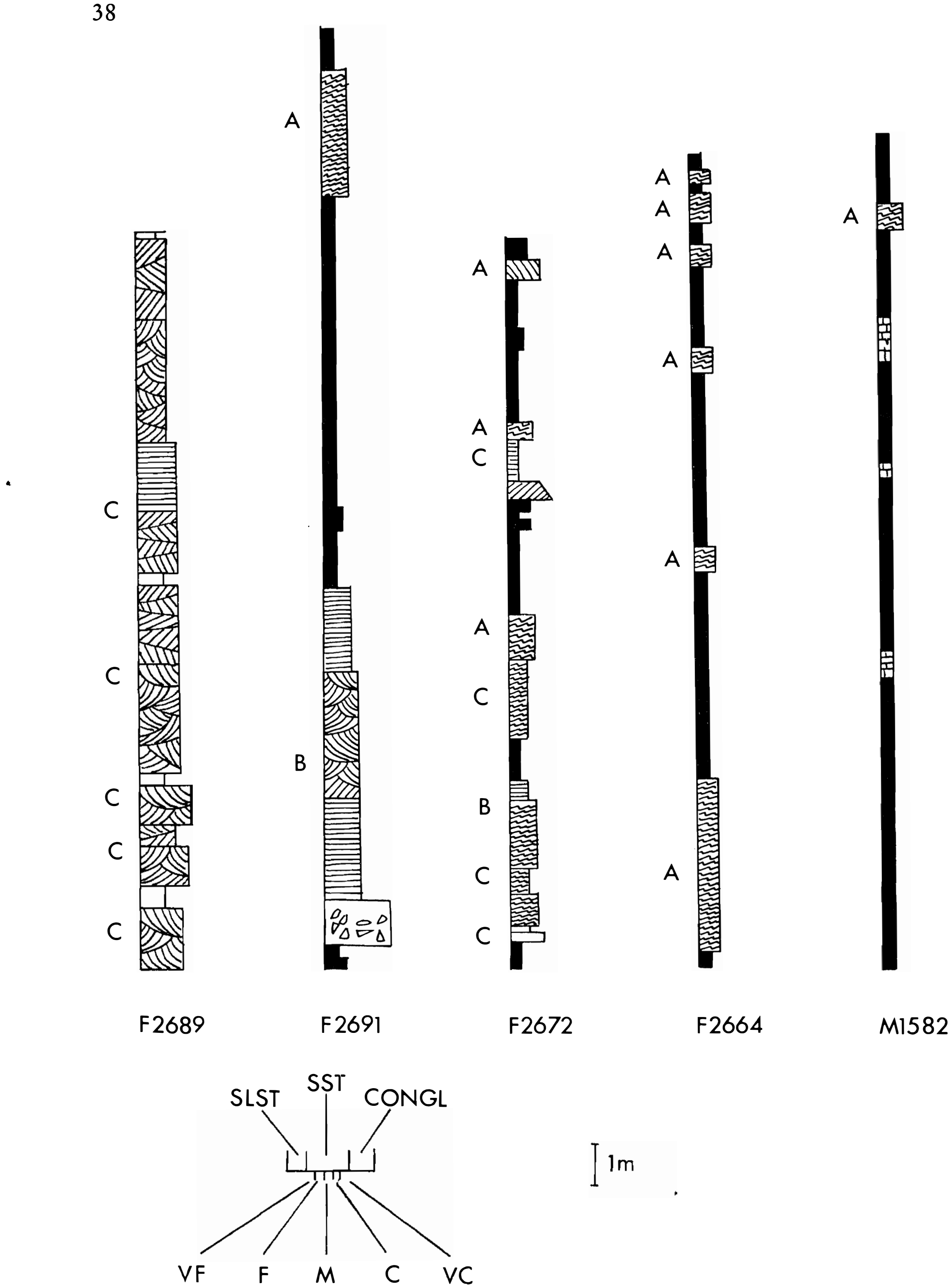

Fig. 22. *Parts of four 'vertical' sedimentological sections, deliberately chosen to illustrate four different patterns of grain-size variation which occur in the Wood Bay Formation. The letters indicate semi-cycles consisting of A) single-set sandstone and siltstone, B) multiple-set sandstone and siltstone, and C) no siltstone.*

load accumulation, to an episode of composite bed-load sedimentation, and back again, must result from important lateral movements of the river channel complex. Steady meander migration within the channel complex will not be enough. In two sections in the Recent deposits of the Murrumbidgee River (SCHUMM 1968b, p. 27), three sand members alternate with silt members. These sand members are quite discrete bodies, and therefore represent distinct episodes of channel location, rather than stages in a continuous and gradual sequence of lateral movements. The Mississippi River meander belt has moved as a whole by distinct steps four times over the last 2,000 years (FISK 1952).

(C) *No siltstone.* In this category the siltstone is either absent altogether, or impersistent. These semi-cycles occur in two main situations.

Firstly they may form the completely sandy sections (e.g. F2689 of Fig. 22) which are typical of areas of bed-load channel deposition. These semi-cycles (C1) are the deposits of the bars of braided rivers. Any tendency to become finer in grain-size upwards, may be due to the mechanism proposed by ALLEN (1970) and described above, or it may result from fluctuations of local flow strength or sediment supply.

Secondly they may occur below other sand semi-cycles in silt-bearing successions (e.g. F2672 of Fig. 22), where they form the lower parts of complex sand units. These complex units could arise in two ways:

(C2) by superposition of two semi-cycles of types A or B, with intervening erosion of silt that had been deposited.

(C3) by complex deposition during an interval of bed-load channel activity. This implies a distinct period of complex channel activity between periods of prolonged suspended-load (overbank or lacustrine) deposition. This is what one of us (MOODY-STUART 1966) had in mind in describing his "low sinuosity" model.

The general sediment pattern of this low sinuosity model is regarded by SCHUMM (1968a, p. 1582) as typical of his mixed-load type. ALLEN (1970) has doubted its existence in nature. We have certainly not found clear published descriptions of Recent deposits of this type formed by bed-load or low-sinuosity rivers. The few descriptions available of bed-load channels with overbank suspended-load deposits, stress that suspended load deposits are usually destroyed by channel movements (SCHUMM 1968a; COLEMAN 1969). However, MOODY-STUART did establish the distinctiveness of the sedimentary association in the Wood Bay Formation. The absence of modern examples may result from a natural tendency to select, for examination, systems which appear to conform with existing knowledge.

Channel depth. – The estimation of depth is probably the most critical and difficult step in our consideration of the alluvial variables. We would stress that any use we make of these alluvial variables will be based on the average properties of a large number of semi-cycles. We are anxious, therefore, to establish that the averages of certain properties are likely to be related to each other, rather than that individual measurements necessarily correspond.

We intend to use the average thickness of the sand members of our semi-cycles, as a measure of average channel flood depth.

We shall now attempt to justify this for the various semi-cycle types described in the previous section:

(A) *single-set sandstone and siltstone.* The amount of bed-load sediment being moved at any one time depends on the stream power (BAGNOLD 1966), which depends on water depth, amongst other variables. All other factors being equal, a thicker sandstone set is likely, therefore, to have been deposited from a deeper water flow.

(B) *multiple-set sandstone and siltstone.* Bank-full (flood) depth is generally equivalent to sand-member thickness, because the sand-member is interpreted as a complete bar deposit (as discussed above).

(C) *no siltstone.* In all types (C1, C2, C3) the semi-cycles are relict fragments of sand-members equivalent to those of types A and B. The thickness of these semi-cycles is therefore a minimum measure of bar, and channel depth.

In general we shall use the thickness of the sand members of our semi-cycles, as statistical measures of mean minimum flood depth.

An independent indicator of water depth is the height of sets of cross-stratification. The heights of large-scale ripples (dunes) with relatively straight crests have been plotted by ALLEN (1968, p. 139) against the depth of the water in which they formed. In his examples the depths varied from 0.5 to 38 m. The ratio of depth of water to height of ripple ranges from 3 to 22, with a mean at 8. ALLEN points out, however, that many of the small and large scale ripples which have strongly sinuous crests form in water only slightly deeper than ripple height. Our cross-bedding sets are the relict fragments of ripples (dunes), some with obviously sinuous crests, and some nearly straight. We feel that it is reasonable to base an assertion about water depth on our statistical observations of cross-bedding set thickness.

The north-western group of sections (LW1 in Fig. 20) have an average sand member thickness of 0.8 m, which we shall use as a measure of mean minimum flood depth. The mean cross-stratification set-thickness for the same sections is 0.2 m (Fig. 23). In contrast the south-eastern groups of sections (UE2 and 3) would have mean minimum flood depths of 2.2 and 2.6 m, and mean cross-stratification set thicknesses of 0.5 m (Fig. 23). This internal consistency encourages us to feel that our assumptions are reasonable.

Friction factor

Hydraulic engineers use a number of different equations to relate mean flow velocity to depth and slope. These equations can be regarded as relating the kinetic energy of the stream flow to the energy loss resulting from frictional drag at the stream boundary. The successful use of these equations depends in practice, on the estimation of the friction or resistance factor, or coefficient. The DARCY-WEISBACH equation is recommended by CARTER and others (1963, p. 99) in their progress report to the American Society of Civil Engineers on friction factors in open channels. They recommend it, in slight preference to the similar CHEZY-MANNING equations, which are rather less widely applicable and less generally used.

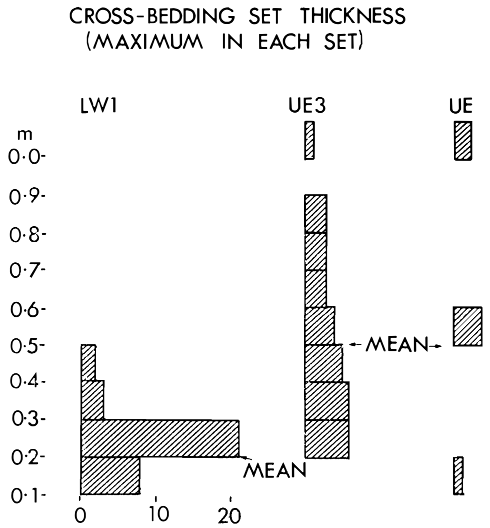

Fig. 23. *Histograms showing the occurrence of cross-bedding sets of various thicknesses in certain groups of localities (positions of groups shown on Fig. 18).*

$$U = \left(\frac{8g}{f}\right)^{\frac{1}{2}} (DS)^{\frac{1}{2}} \quad \text{Darcy-Weisbach,} \qquad (1)$$

where f = friction factor or resistance coefficient, U = mean velocity, D = depth, S = slope.

In alluvial channels, frictional losses depend on (a) grain-size of the sediment, (b) nature and size of bed forms (e.g. ripples, plane beds etc.), (c) form of channels (sinuosity, size etc.).

It remains to consider the actual values of the friction factor or resistance coefficient (f). Bagnold (1966, Table 1) has published a list of observations on certain rivers, mainly in south-western North America. These observations include mean velocity, depth and slope, and it is therefore possible, using the above equation, to calculate 'f' for each set of observations. Fig. 24 shows the variation of values of "f". It is found that 95% of the 146 values lie between .01 and .15, and that the mode is .04. These values are therefore used in this investigation.

Many of the rivers which make up Bagnold's list are large. They are probably larger than the rivers responsible for much ancient river sedimentation. At the other extreme of size are the laboratory flume runs reviewed by Guy, Simons and Richardson (1966). Of the 8′ runs involving 0.19 mm, 0.45 mm, and 0.93 mm sand, 100% of the "f" determinations lie within the range selected here (0.01 to 0.15). Allen (1970, Fig. 14, p. 316) uses the same flume data to show a distinction between the friction factor for plane beds and antidunes (f = 0.02), and for ripples and dunes (large-scale ripples) (f = 0.08). In our Spitsbergen regional work, cross-strata formed in ripples or dunes are much the commonest structures. As we shall show below, the assumption of a single modal friction factor does not seem to be a serious source of error in our sort of work.

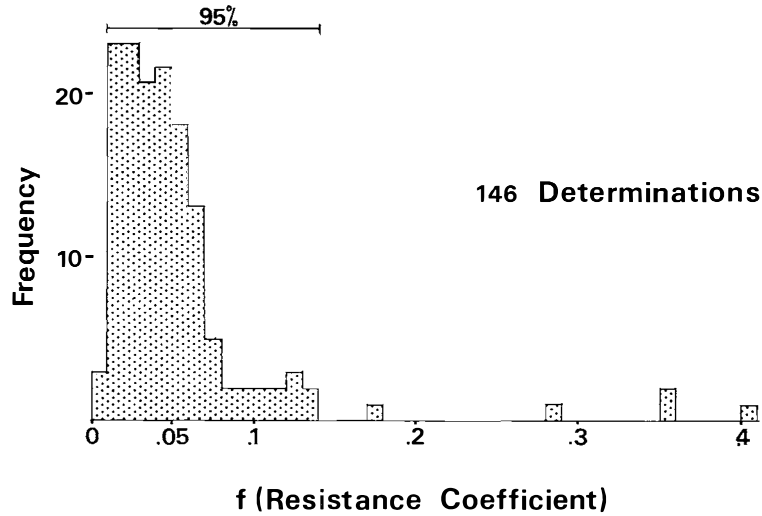

Fig. 24. *Histograms showing the occurrence of different values of the Resistance coefficient (f) for the 146 present-day measurements of rivers listed by* BAGNOLD *(1966, Table 1).*

Strength of River Flow

Arbitrary choice of index. – HJULSTRÖM's well-known diagram (1935, p. 298) used water flow velocity as an index of the erosion, transportation and deposition of sediment of different grain-sizes. Some such index of stream strength seems a valuable measure of the processes which do much to control sediment movement. Many other measures of stream strength have been used. For instance, in addition to velocity, WILLIAMS (1967, p. 7) gives four measures (stream power, shear stress using depth, shear stress using hydraulic radius, regime theory bed factor). He comments, "No single measure of flow strength has yet gained wide-spread acceptance".

Since that was written, the work of BAGNOLD (1966) has done much to establish the use of stream power (defined below) as an index of stream strength, for portraying the ability of a stream to move sediment. BAGNOLD regards the transporting of sediment as a form of work carried out by the stream, and its power as a measure of the rate at which this work can be performed. Having developed this idea theoretically, BAGNOLD supported it by presenting detailed measurements made in laboratory flumes and natural rivers. We have decided to use stream power as one index of river strength.

The definition of stream-power involves shear stress, which itself involves slope, as will be described below. KENNEDY (1971, p. 132) points out that "sediment transport rate and flow velocity are not particularly sensitive to slope and shear stress". In addition, single values of slope sometimes correspond to three different values of mean velocity, even with constant discharge (because of variation in bed-form development). "Depth-discharge predictors which use slope or any quantity including slope as an independent variable, are inherently inferior in at least two respects (uniqueness and sensitivity) to those which treat S as a dependent variable". In our work, we use slope, depth and a highly generalised friction factor to estimate a stream-power for modern rivers. This use is clearly open to criticism on the grounds put forward by KENNEDY. However we feel that our use of stream power is justifiable over the range of river strengths which concerns us, which is much larger than that being considered by KENNEDY. Our interest is the distinction between silt, sand and pebble movement. We also have an interest in utilising stream power as an index, because of its value in measuring the rate of river work. We shall, however, also quote mean velocity as a strength index.

Relationship of velocity and power to depth and slope. – The relationship between velocity and depth and slope has been considered above.

The power of stream flow (ω), per unit area of stream bed, is defined as the product of the mean velocity (U), and the bed shear stress per unit area (τ).

$$\omega = U\tau \tag{2}$$

The shear stress (τ) is then defined as follows, where p = density, g = acceleration due to gravity, D = depth of water, S = slope of water:

$$\tau = pg\, DS \tag{3}$$

This stress is the downslope component of the weight of the water. The relationship is approximately true only if 1) the slope, S, is a low angle, 2) the flow is uniform, 3) the depth, D, is small compared with the width (WILLIAMS 1970).

Combining equations 1, 2, 3

$$\omega = pg\,(DS)^{3/2}\left(\frac{(8g)}{f}\right)^{\frac{1}{2}} \tag{4}$$

We assume that p and g are constant and fix f at 0.04 (see above) in our particular situation. On plots of log D against log S therefore a series of straight and parallel lines will represent different values of mean velocity and power. (Figs. 25, 26).

$$\log D = \text{constant} - \log S \tag{5}$$

Velocity and power increase with increases of depth or of slope. To achieve fixed velocity or fixed power, depth and slope must behave inversely.

Sand bed-forms as indicators of local stream strength. – BAGNOLD (1966) and GUY, SIMONS and RICHARDSON (1966) showed that a systematic change of bed forms occurs in the sand of experimental flumes when the stream power and the transport rate increase. This relationship has been examined in more detail by ALLEN (1968, Fig. 6.9; 1969 Figs. 1 and 2; 1970, Fig. 2.6) who particularly concentrated on the variation of bed forms which is found with differing mean grain-size. He used the experimental data of GUY, SIMONS and RICHARDSON (1966), and WILLIAMS (1967).

We have used these same data, and some provided by LAURSEN (1958), to plot stability ranges of different bed forms against power and mean grain-size, in S. I.

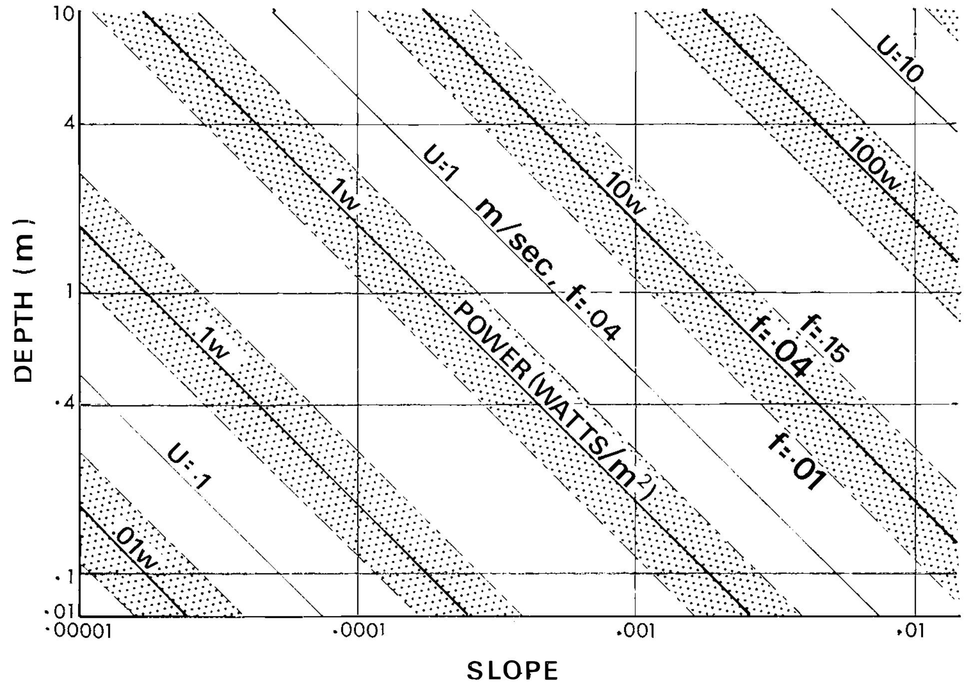

Fig. 25. *Plot of flow depth against slope, showing a line of equal mean velocity* ($U = 1$ *m/sec, where Resistance coefficient* (f) = *0.4), and lines of equal power with Resistance coefficient* (f) = *0.01, 0.04 and 0.15. This plot is based on equations 1 and 4.*

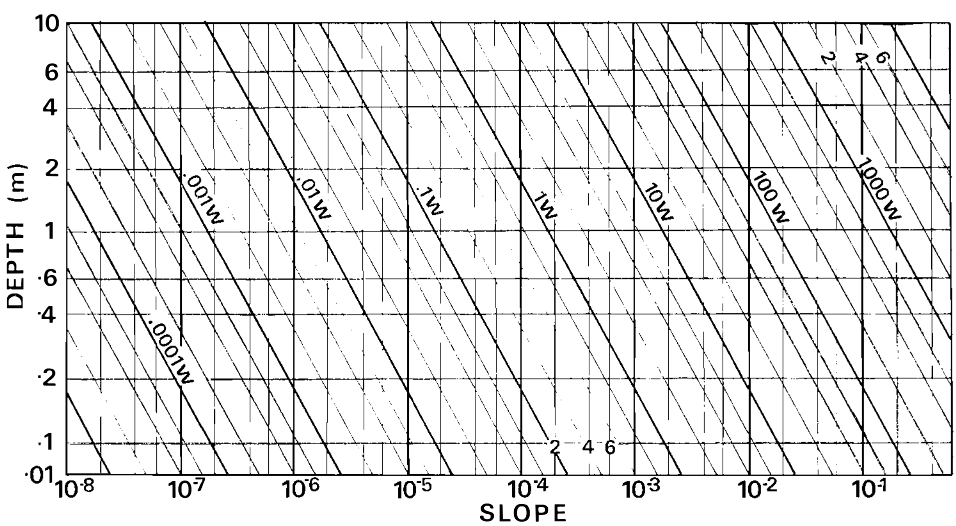

Fig. 26. *Plot of flow depth against slope, showing lines of equal stream power (watts). This plot is based on equation 4, assuming Resistance coefficient* (f) = *0.04.*

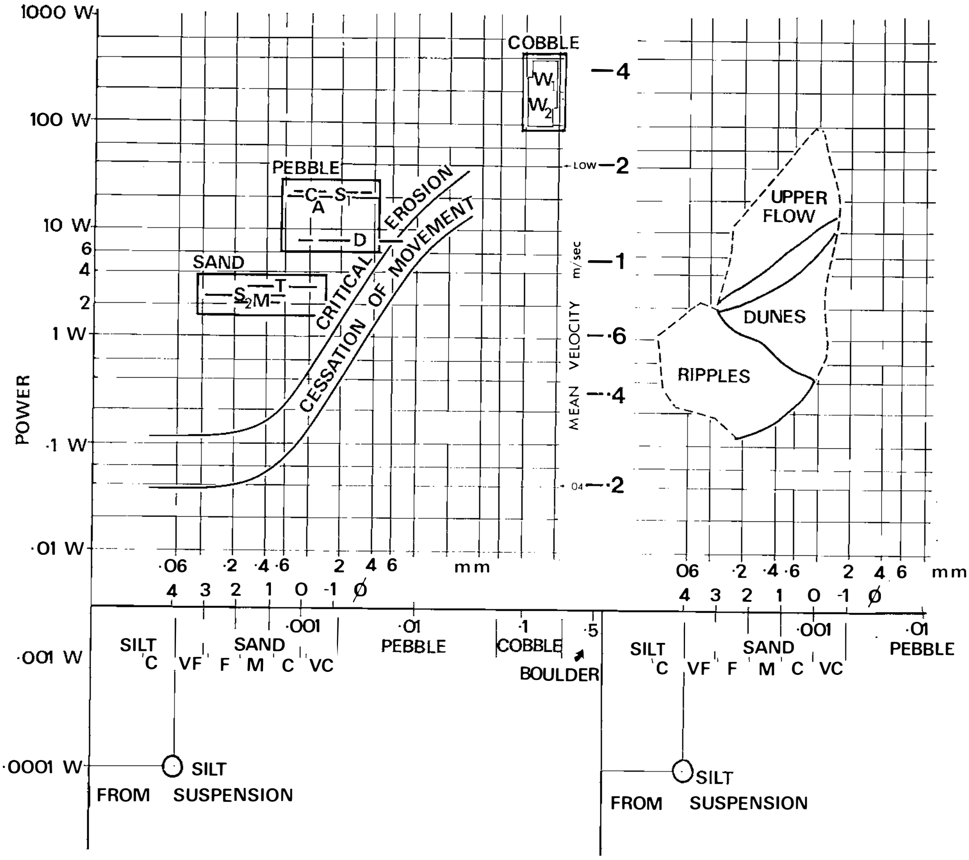

Fig. 27. *Plot of grain-size against stream power and mean velocity, showing* 1) *cessation of movement and critical movement curves* (Sundborg), 2) *alluvial grain-size* (*flood power data from some present-day rivers* (*Table 1*), 3) *grain-size power fields for sand, pebble and cobble rivers and* 4) *the fields of occurence of the most important sand bed-forms* (Guy, Simons *and* Richardson, *1966*).

units (Fig. 27). This plot shows a clear sequence from ripples (or lower-phase flat beds) to dunes (not always represented) to upper phase flat beds, with increasing power. Allen (1970) used variations in power and in grain-size as variables controlling the formation of one or other of the bed forms.

There are limitations to this method. Firstly, when the water is shallow, the general relationship between sand bed-load transport and power will break down completely. Bagnold suggested (1966, p. 19) that this will occur with depths of 2 to 3 centimetres. Secondly, the actual measurements reported by Guy, Simons and Richardson (1966) may refer to conditions rather different from those in an aggrading stream. Equilibrium, rather than aggradation, was the experimental situation, and the power was generally raised to the equilibrium level, rather than dropped to it after a flood peak.

Finally we know so little about the significance, in terms of flood stage (or depth), of individual sets, that we prefer a method which uses a feature of a complete sand member to indicate the flood power of the river. This we shall consider below.

Grain-size generally as an indicator of general stream strength. – It is common knowledge that boulder beds are deposited by stronger streams than silt beds. It is our object in this section to consider this quantitatively.

For over 150 years, hydraulic engineers have been using empirical, and partly empirical, relationships between grain-size and velocity of flow (LELIAVSKY 1966, p. 34). Their work has often been concerned with problems of scour, so that "pick up" velocities ("critical erosion velocities") are usually quoted. However velocities for the cessation of sediment movement are also important in the sedimentation process. For any particular grain-size, cessation of sediment movement occurs at about 2/3 of the critical erosion velocity (HJULSTRÖM 1935, p. 320; MENARD 1950, p. 151; SUNDBORG 1956, p. 181).

For our purposes we have used the plots of velocity against grain-size published by SUNDBORG (1967, p. 337). He quoted velocities 1 m above the sediment surface, and we have converted these to mean velocities by multiplying them by 0.8 (HJULSTRÖM 1935, p. 294; SUNDBORG 1956, p. 175). We have then converted these mean velocities to powers using a combination of equations 1, 2 and 3:

$$\omega = \frac{U^3}{8} pf \qquad (6)$$

where ω = power, U = mean velocity, p = density, and f = friction factor.

This becomes $\omega = 5U^3$, in S. I. units, assuming standard friction factor, f = 0.04.

Curves for critical erosion, and for cessation of movement, are then plotted on Fig. 27, against axes of mean velocity, power and grain-size. Assuming that clasts of all sizes of material were available in the source areas, we would expect the flood velocities and powers of rivers to be reflected in the grain-size of their alluvium.

We have already pointed out that many local processes may influence the grain-size of particular beds. These include 1) processes influencing the grain-size of material supplied to the alluvial area, 2) processes of local hydrodynamic sorting, 3) processes causing local fluctuations of velocity or power, due to bed forms, channel shape etc.

We hope to avoid some of these difficult problems by basing our method on a general range of grain size for our alluvial systems, rather than the actual grain-size of local features. On Fig. 27 we have plotted ranges of grain size against power or mean velocity estimates for a number of modern rivers (Table 1), from the fine and medium sand of the lower part of the Mississippi to the cobbles of the White River. The estimate of river strength are flood estimates using the relationships to depth and slope (equations 1 and 4), and a constant friction factor of f = 0.04. Estimates of flood depth, particularly, are often difficult to make from the published descriptions, but the possible errors are not serious for our purposes at this stage.

On the basis of the grouping of these data, we distinguish:

1) *sand rivers,* with a median flood power of about 2W

2) *pebble rivers,* with a median flood power of about 15W

3) *cobble rivers,* with a median flood power of about 200W

Table 1
Grain-size and flood power of some present day rivers

Mississippi, U.S.A. (M) (Frazier and Osanik 1961; Kolb 1963; Leopold, Wolman and Miller 1964; Moore 1970), *alluvial tract.* fine-medium sand; flood depth, 30–34 m; slope 8 cm/km. Power, therefore 2W.
Slims River, Canada (S) (Fahnestock 1969)
Intermediate (S_1), medium gravel, sand; flood depth, 1.3 m; slope 100–330 cm/km. Power, therefore, 20W.
Downstream (S_2), sand, silt; flood depth, 1 m; slope 30 cm/km. Power, therefore, 2W.
Tana River, Norway (T) (Collinson 1970), medium to coarse sand; flood depth (based on our calculations from discharge figures), 2.5 m.; slope 20 cm/km. Power, therefore, 3W.
Amite (A), Louisiana, U.S.A. (McGowen and Garner 1970), coarse sand and pebble gravel; flood depth, 7 m; slope 60 cm/km. Power, therefore, 20W.
Colorado (C), Texas, U.S.A. (McGowen and Garner 1970), coarse sand and pebble gravel; flood depth, 11.5 m; slope 30 cm/km. Power, therefore, 20 W.
White River (W), Washington, U.S.A. (Fahnestock 1963, p. 26 etc.). Above moraine (W_1) cobbles (.12 – .18 m), flood depth about 0.8 m; slope 3000 – 8000 cm/km. Power, therefore, 150 – 400W. Below moraine (W_2), cobbles (.1 – .16 m), flood depth, about 0.8 m; slope 2000 – 6000 cm/km. Power, therefore, 100 – 300W.
Donjek River (D), Canada. (Williams and Rust 1969; Rust personal communication). Gravel most abundant, sand; flood depth about 2.8 m (our calculation using flood discharge estimate); slope 60 cm/km. Power, therefore, 8W.

All the Wood Bay rivers deposited alluvium of very-fine to coarse sand. They were all "sand rivers", and as the curves of Fig. 27 show, changes within the sand grade appear to make little difference to the power ratings. We therefore suggest that all the Wood Bay rivers were characterised by flood velocities of about 0.7 m/sec., and flood powers of about 2W. The presence of conglomerates would greatly extend the range of flow-strengths implied.

Slope

Slope, the last of our dependent variables, is estimated by using the depth-slope-power relationships of equation 4, which we have plotted graphically in Figs. 25 and 26.

Assuming a general friction factor of 0.04, and a general flood power of 2W, then slope will depend on fluctuations of flood-depth. The local slope of the channels is converted to a more regional slope by multiplying it by the sinuosity.

RIVER SYSTEMS AND INDEPENDENT VARIABLES

We use the term "river system" to describe the entire catchment areas of a river or group of rivers. A river system is distinguished from other river systems on the basis of similarity of the palaeocurrents and sedimentation characteristics of its alluvial parts. The "alluvial part" of any river system receives its detritus from the "source part".

Before interpreting the channel deposits of the various river systems, we shall mention a number of features which, in various ways, point to a complementary environment of sedimentation. This is one in which predominant sedimentation was from suspension either in flood lakes, or longer-lived lakes.

The Clay flats

Siltstone sequences. – Much of the Wood Bay Formation consists of siltstone, generally red in colour. Most of this siltstone contains fossil burrows and calcareous concretions (FRIEND and MOODY-STUART 1970). It is generally, however, devoid of ripples or cross-stratification, and we conclude that it was deposited from suspension.

In most areas these red siltstones are interbedded with sand members, which we have interpreted as river deposits, and we have already analysed the great variation in the thickness and separation of these (Section 4). In some other areas (Fig. 28), sand members are either absent or represented by one individual, so that thicknesses of up to 35 m are composed entirely of siltstone.

It is difficult to decide whether these siltstones resulted from deposition in short-lived lakes (existing for days or months), or whether they accumulated in longer-lived lakes which existed for years, or even thousands of years. However the fact that these sediments are red implies that any contained organic material was oxidised (FRIEND 1966) while they stood above water table. The carbonate concretions similarly imply periodic drying out (FRIEND and MOODY-STUART 1970). On these grounds we suggest that the lakes were mainly short-lived.

KRINSLEY (1970) has recently described the geomorphology and conditions of deposition of the playas of Iran. "Clay flats" form about 35% of the total playa area in Iran, being second only in abundance to "salt crust" areas where the salt is due directly to the reprecipitation of locally outcropping Miocene evaporites. Many of the playas are entirely occupied by clay-flats. A clay flat playa "may be the bed of a former lake, or it may be an interior sump that is seasonally inundated by a thin sheet in which fine sediment load from the surrounding alluvial fan streams is transported" (KRINSLEY 1970, p. 281). "Lake" Stefanie of the African Rift Valley seems to be similar to these playas in its clay-flat floor and intermittent flooding (GROVE and GOUDIE 1971).

Carbonate members. – We have recently described, in a separate publication, the bedded carbonates of the Wood Bay Formation (FRIEND and MOODY-STUART 1970). The sediments are marlstones, contain up to 68% carbonate, and are in some cases, dolomitic. Individual units are generally between 1 and 3 m thick, and one case is known of a unit extending 9 km (Fig. 29). Several of these units occur in each stratigraphical "carbonate member" (Fig. 4), as defined by FRIEND, HEINTZ and MOODY-STUART (1966).

We think that each of these units formed in a lake. Often the lakes contained large numbers of ostracods and charophyte algae, and carbonate precipitation may have been stimulated by the algae. To allow the steady accumulation of this thickness and extent of carbonate, the lakes must have been relatively long-lived, perhaps lasting for thousands of years. The lakes were probably about the same depth and extent as the individual marlstone units (metres, by, in one case, 9 km).

KRINSLEY (1970) described the lakes which occupy parts of playa surfaces in present-day Iran. Their position is determined by the proximity of a source of surface water discharge, and by ground water conditions. In Spitsbergen the

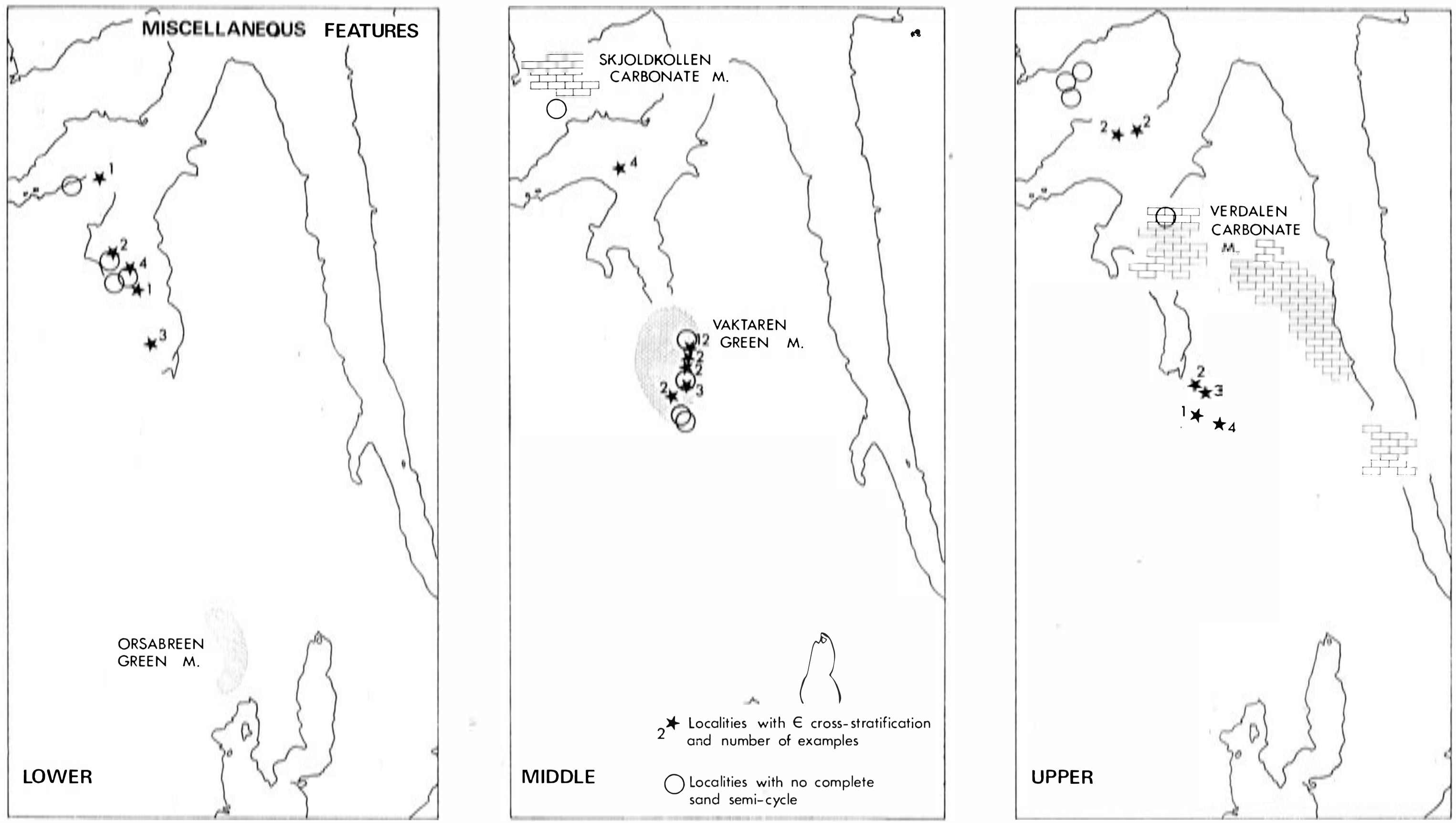

Fig. 28. *Standard map area (outlined on Fig. 1), for the three informal time-rock units, showing certain miscellaneous features: the extent of carbonate and green members, the occurrence of sections with no sandstone units, the occurrence of sections with epsilon cross-stratification.*

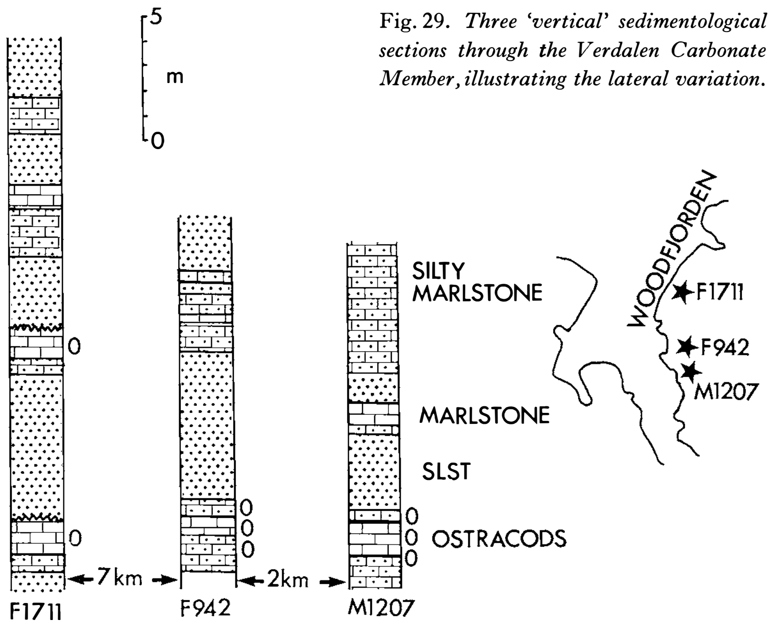

Fig. 29. *Three 'vertical' sedimentological sections through the Verdalen Carbonate Member, illustrating the lateral variation.*

carbonate members formed in periods of recurrent lake existence and they therefore represent the local recurrence of particular hydrological conditions.

Green members. – In two different areas (Fig. 28) and at different stratigraphical levels (FRIEND, HEINTZ and MOODY-STUART 1966), zones of unusually persistent green colour occur in the Wood Bay Formation. These zones are lenticular in form, up to 50 m thick, and extend for minimum distances of 15 and 20 km. A coloured photograph of the Orsabreen Green Member is reproduced on the first page of a popular article by one of us (FRIEND 1969, p. 689).

In distant view, the green members are easy to confuse with the carbonate members described above, and this has caused some trouble in working out the Wood Bay stratigraphy. However they are not richer in carbonate, or poorer in sandstone and siltstone than the surrounding red Wood Bay successions. They contain fining-upwards sand and silt semi-cycles, which are only unusual in their lack of redness. They appear to have similar permeability to the normal red successions. These green members differ from the Austfjorden Sandstone Member, which although uniformly grey or green, consists of sand only.

We interpret the green members as the result of general reduction of iron oxides in the sediment, due to unusually high level of the water table (FRIEND 1966). They appear to reflect persistent "boggyness" in certain areas. This is the characteristic of the "wet zones" where Iranian playas meet the neighbouring alluvial fans (KRINSLEY 1970, p. 264).

Epsilon cross-stratification. – This name was applied by ALLEN (1963) to cross-stratification formed by lateral deposition on the sloping surface of a channel bar. The feature which is diagnostic of this structure is the presence of current structures which indicate that local flow was dominantly along the strike of the major foresets.

MOODY-STUART (1966) has illustrated and described many occurrences of this structure in the Wood Bay Formation. He used these occurrences (Fig. 28), and a number of other features (MOODY-STUART 1966, p. 1116) to distinguish high sinuosity streams from low sinuosity streams. The high sinuosity streams meandered through sediment which was generally finer-grained and redder (p. 1114), and they appear to have been down-slope from the low-sinuosity streams and at the edge of the clay flat areas. KRINSLEY (1970) gave numerous examples of channels meandering across the clay flats of the Iranian playas.

Pattern of river systems

Fig. 30 represents graphically the alluvial systems which we have recognised. Table 2 gives some values for the dependent variables, grain-size, sand member thickness, channel type and current direction. Our methods of estimating these have been discussed above. We have made no attempt to give the range of these values. In most cases we have given average, or characteristic, values, although in one case (where the modes are particularly distinct) we have given two modal values.

One of the clearest features of this analysis is the existence, in all three periods, of a distinctive eastern system. This was characterised by bed-load channels, averaging at high stage, at least 2 to 3 m deep, 120 to 170 m wide, and of low sinuosity. These were of braided appearance at low stage. The regional slope was gentle (15 cm/km), and towards the NNW. In "Upper" times the bed-load channels poured downstream into mixed-or-suspended load channels of high sinuosity, which, in turn, poured into clay flats. Major channels appear to have divided, and then petered out altogether. This is a common situation in 'internal' drainage basins, and in arid areas.

The western system provides a complete contrast with this. This system was

Table 2

	Lower				Middle		Upper	
	E	C	W	E	C	W	E	C
Maximum grain-size (sand-grade)	F–C	F	F–C	F–C	F–C	F–C	F–C	F–M
Sand member thickness = depth	2.9	1.2	0.7	2.2	0.6	0.25 (3.25)	2.0	1.3
Dominant channel type	B	M–S	M–S	B (M-S)	M–S	M–S	B (M–S)	M–S
Width/depth	60	10	10	60–10	10	10	60–10	10
Width (m)	170	12	7	130–22	6	2.5 (32.5)	120–20	13
Sinuosity	1	2	2	1–2	2	2	1–2	2
Flood mean velocity (m/sec)	0.7	0.7	0.7	0.7	0.7	0.7 (0.7)	0.7	0.7
Flood power (W)	2	2	2	2	2	2	2	2
Slope – Channel ($x10^5$, cm/km)	15	30	50	18	50	160 (10)	20	30
– Floodplain (»)	15	60	100	18–32	100	320 (20)	40	60
– Direction	NNW	N	ENE	NNW	NNE	ENE	NNW	N, NNE

B = bed load; M–S = mixed-or-suspended load.

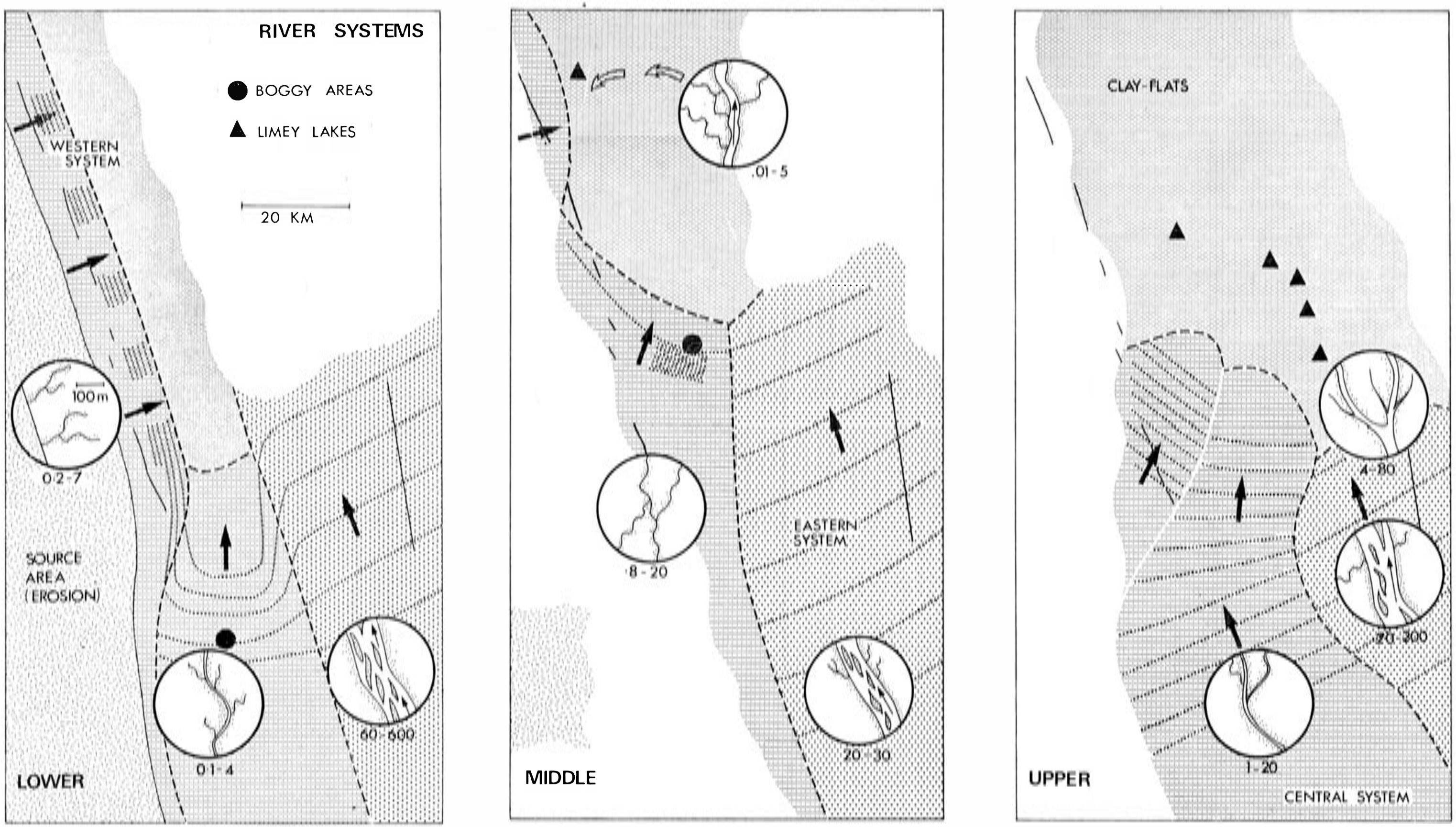

Fig. 30. *Standard map area (outlined on Fig. 1), for the three informal time-rock units, summarising the properties (Table 2) of the three river systems and the clay-flats. Inside the large circles are diagrammatic impressions of the sizes and patterns of the river channels, and below them are estimates of mean discharge and mean annual flood (m^3/sec). Diagrammatic contours on the alluvial surfaces are at 1 m intervals.*

clearly defined in "Lower" times, and present in "Middle" times. There is no evidence for, or against, its presence in "Upper" times. In "Lower" times channels with flood depths of about 0.7 m, were of mixed-or-suspended load type. They were highly sinuous, and about 7 m wide. They deposited sediment on a relatively steep slope (100 cm/km), which faced ENE, and extended from a source only a short distance away. This system did not extend more than 10 km into the basin, where it gave way to an area of clay flats. We presume that the channels split into distributaries and then petered out altogether. In "Middle" times, the western system is similar except that there is additional evidence for the presence of a much larger (3 m deep) highly sinuous river, into which the smaller, and more typical channels may have drained.

The slopes deduced here seem, at first sight, surprising. The smaller western system rivers were forming a steeper slope than the larger eastern system rivers. However accurate contours on the present-day Indo-Gangetic alluvial-plain, show a similar effect (GEDDES 1960, Fig. 1).

The source areas for this western system seem to have been quite small in area. The sand members are laterally impersistent, and they tend to be concentrated in certain sections (Fig. 21), suggesting that they formed in the channels which emerged from nearby, relatively long-lived, valley mouths. The size of the channels also suggests a small catchment area, and we suggest a typical area of 5 km by 5 km (25 km^2) for a single source area.

In the south-west of our area, there was a river system which we shall call the central system. This is characterised by a northerly or north-north-easterly palaeocurrent direction, and a high silt to sand content.

It had a similar alluvial slope to the eastern system, but sedimentation more like the western system. Mixed-or-suspended load channels, varying from 0.6 to 1.3 m deep, and 6 to 13 m wide, were of relatively high sinuosity. The alluvial slope varied from 60 to 100 cm/km and was relatively consistently N or NNE in direction. The two green, or boggy, members were formed in this system, and it may be that the poor drainage resulted from the proximity of the converging western and eastern systems.

Independent variables controlling the river systems

Having estimated values of the dependent variables, we can now consider the first independent variable, water discharge.

Water discharge. – SCHUMM (1969, p. 256) has presented empirical relationships between depth of channel, channel perimeter silt content (itself related to channel type) and either mean discharge (annual) or mean annual flood. He used 36 sets of data from the western U.S.A. and Australia. We plot these relationships (Fig. 31), using S. I. units and some arbitrarily chosen values of the perimeter silt content (M). We take M = 5% as characteristic of bed-load channels, and the characteristic for mixed-or-suspended load channels as somewhere between M = 12% and M = 25%.

In Table 3, we have listed our estimates of discharge for the average rivers

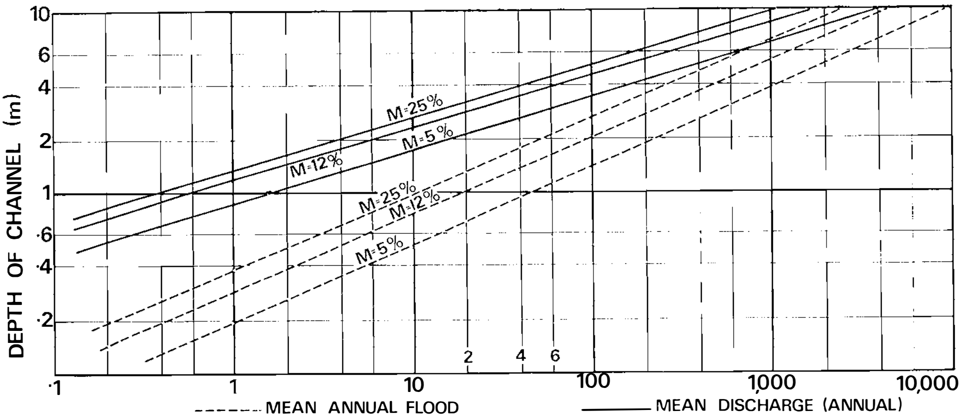

Fig. 31. *Plot of river channel depth and mean annual flood and mean discharge (m^3/sec) based on the relationship of* SCHUMM*(1969 p.256). Different values of* SCHUMM's *M (channel perimeter silt content) are plotted.*

represented in the various systems of the Wood Bay Formation. The mean annual floods of these average rivers of the eastern system were about 100 times greater than those of the western systems.

LEOPOLD, WOLMAN and MILLER (1964, p. 251) suggested an approximate relationship between flood discharge and drainage area.

$$Q \propto A^{0.75}$$

Although this was presented as a means of estimating discharge at different points in one river system, we shall use it to compare the drainage areas of the western and eastern systems. We assume therefore, for the purposes of this calculation, that rainfall in the two areas was similar

$$\frac{Q_1}{Q_2} = \left(\frac{A_1}{A_2}\right)^{0.75}$$

On the basis of this equation, an eastern source area was about 400 times greater than a western source area. If we use 25 km² as an estimate of the area

Table 3
Average River Discharges

	Depth (m)	M %	Mean discharge m^3/sec	Mean flood m^3/sec
Eastern	2.9	5	60	600
	2.2	5	20	300
	2.0	5	20	200
Central	0.6	12–25	0.1	4
	1.2	12–25	0.8	20
	1.3	12–25	1	20
Western	0.7	12–25	0.2	7
	0.25	12–25	0.07	0.5

upstream of one of the rivers of the western system, we arrive at an estimate of 10,000 km² (100 km by 100 km) for the source area of one of the eastern system rivers. A difference of this sort is the basis of Fig. 33.

In order to test the importance of our assumption of uniform rainfall, we shall also use a 'contrasting rainfall model'. We shall define this model as one in which the eastern system had four times as much rainfall as the western one. The eastern system rivers could, in this model, drain a source area a quarter the size of the one calculated above, i.e. 2,500 km² (Fig. 33).

On Fig. 30, our graphical representations of the channel dimensions and plans, include estimates of meander wavelength. These use SCHUMM's (1969, p. 258) equations relating discharge to meander wavelength (Fig. 32), and channel perimeter silt content.

Nature of sediment discharge. – Our brief discussion of the total sediment load of the Wood Bay Formation rivers, will be postponed to our last section, "Tectonic Implications".

In this section we are concerned to compare and contrast properties of the river systems which appear to reflect differences in sediment discharge. The differences between bed-load and mixed-or-suspended load types of river channels seem to depend largely on their sediment loads (SCHUMM 1963; 1968b). Bed-load channels are thought to have more than 11% sand or gravel in their sediment load (SCHUMM

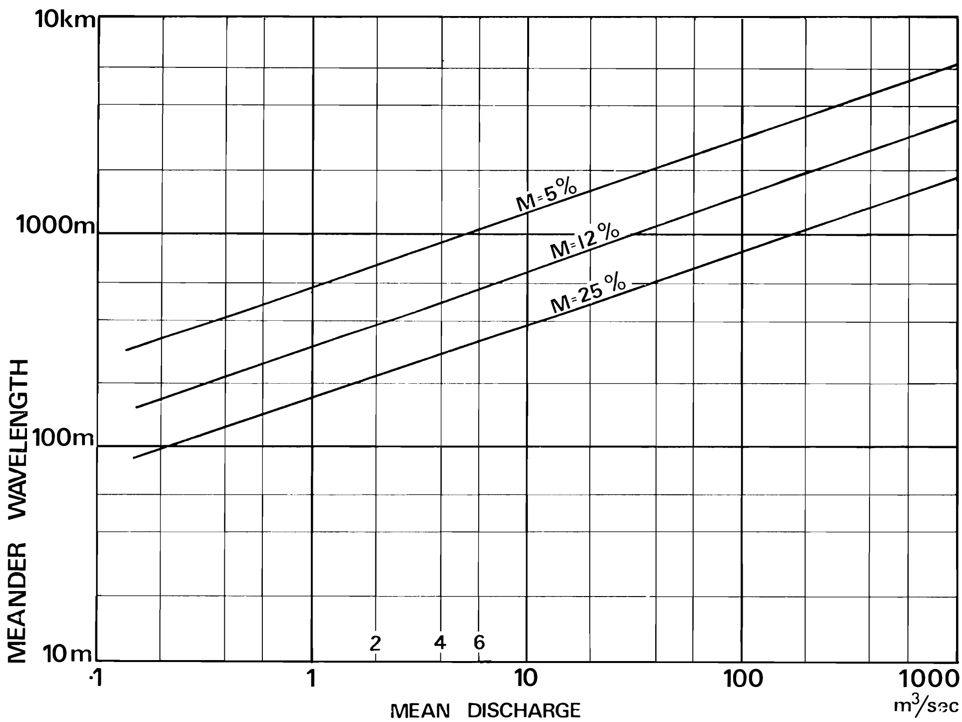

Fig. 32. *Plot of meander wavelength and mean discharge using relationship quoted by* SCHUMM (*1969, p. 258*). *M is* SCHUMM's *channel perimeter silt content index.*

1968b, p. 40). In the discussion which follows we shall direct our remarks at the bed-load channels and the bed-load deposits of the Wood Bay eastern system, with the implication that they are distinct from the mixed-or-suspended load channels and the deposits of the western and central systems.

Sediment load appears to be more critical in determining whether a channel is of bed-load type, than any feature of the water discharge pattern. The Ganges and Brahmaputra Rivers have remarkably similar curves of annual discharge variation (Coleman 1969, p. 153), particularly in terms of the size of peak floods. Yet the Brahmaputra is braided (and presumably has bed-load channels), whereas the Ganges is primarily a meandering (presumably mixed-or-suspended load channel) River. The reason for this is that "the Brahmaputra is slightly more heavily charged with sediment" (Coleman 1969, p. 155). This last statement is supported by measurements of suspended load, but Coleman's description of the size and velocity of bed forms makes it difficult to believe that bed-load is not also distinctly greater in the Brahmaputra River.

Schumm (1969, p. 266) cited examples of river channels which have changed from one type to another in historical times. Rivers of the Sacramento Valley in California (Gilbert 1917) became bed-load rivers, when sand and gravel were flushed into their head-waters during hydraulic mining for gold. There was no significant change in overall water discharge. Schumm (1969, p. 266) gives other examples of channel type changes, in both directions, and these usually occurred with changes in the size of peak floods, due to alteration by man of the vegetation or the flow of water. It seems likely, however, that, in all these cases, there was a change in the sediment discharge too.

The distinctive bed-load properties of the eastern system were maintained throughout Wood Bay times (say 20 million years). We are only concerned, therefore, with factors capable of maintaining a long-term (20 million year) difference. Source area factors likely to control sediment yield have been discussed by Miller and Piest (1970).

Sediment yield from the perimeters of the channels of the source area would have been influenced by:

1) water depth and channel slope
2) bed-rock properties

As far as 1) is concerned, we have established above that, in the Spitsbergen alluvial areas, water mean velocity and power were similar. This suggests that they were also similar in the lower parts of the source areas.

Sediment yield by sheet and rill erosion of the areas between the channels would have been influenced by:

3) climate
4) amount of vegetation
5) bed-rock properties
6) length and gradient of slopes

As far as 3) is concerned, we shall at this stage assume quite arbitrarily that there was no climatic difference between the alluvial systems. The consequences of this assumption will however be considered further in a later section.

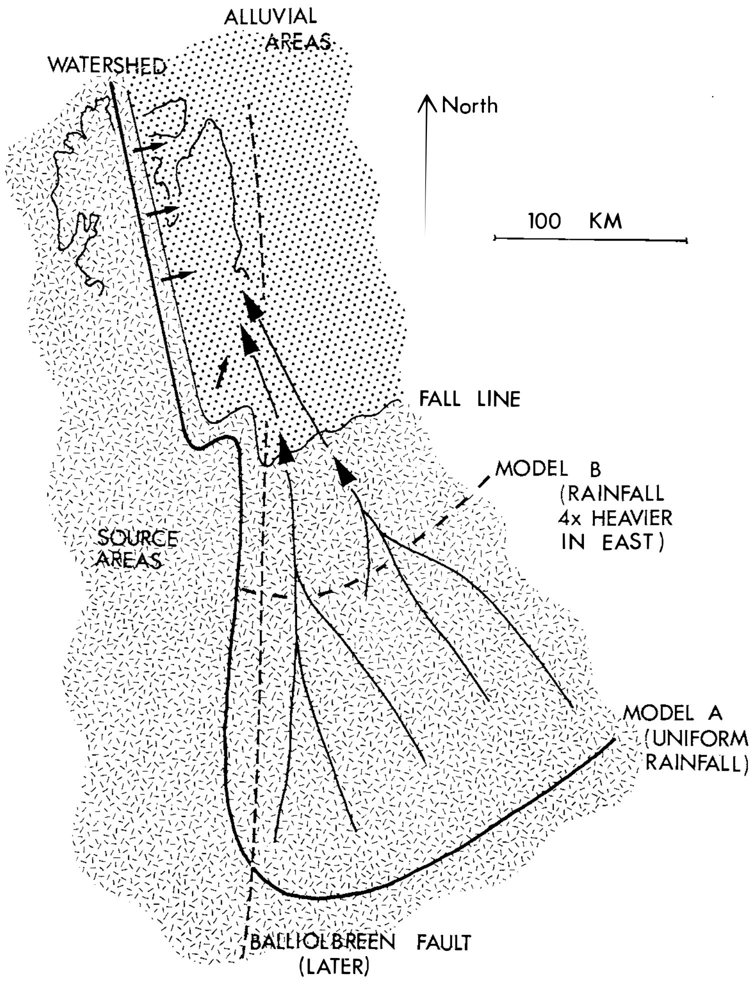

Fig. 33. *Map showing suggested source areas of the sediments of the Wood Bay Formation.*

As far as 4) is concerned, we consider that, in the Devonian, all the systems would have lacked sediment-binding vegetation (SCHUMM 1968a). Vegetation would therefore not have been responsible for differentation between the systems.

It does seem possible that variation of bed rock properties (factors 2 and 5, above) would have been responsible for the differences in bed-load. The eastern source areas may have contained more exposures of bed-rock which weathered to yield sand-grade detritus. The high proportion of orthoclase delivered in Lower and Middle times to the eastern system is evidence for more exposures of migmatites and other feldspathic crystalline rocks. These may have contrasted with the limestones, pelites and quartzites not only in their ability to produce orthoclase, but also in their ability to produce sand-grade detritus.

Finally (factor 6, above), it is quite possible that differences in relief were associated with differences in the length and gradient of slopes, and these may have altered the bed-load yield of the rivers.

Tectonic implications

THE OVERALL PATTERN

The presence of a thick (about 3 km) pile of sediment, involving no major vertical change of sedimentary environment, is evidence in itself for crustal downwarp of the area of accumulation, relative to uplifting source areas. Our work has defined the size and shape of the source areas (Fig. 33), and thus provided information about the movement of this part of the earth's crust during Wood Bay times.

Although this information concerns relative movements of source and basin, we can relate the basin surface approximately to sea-level. In the immediately succeeding Grey Hoek Formation, fossil bivalves provide evidence of marginal marine conditions (FRIEND 1961), and we therefore conclude that the Wood Bay alluvial surface was only slightly above sea-level.

There is some evidence from gravity survey (P. MATON, personal communication) that the Devonian sediments thin sharply north of the northernmost Wood Bay Formation outcrops. We shall therefore make the assumption that all the sediment from the Wood Bay source areas was trapped and deposited in the area defined approximately by the present outcrops.

SOURCE AREA DENUDATION, UPLIFT AND DOWNWARP

Rates of denudation for particular present-day drainage basins have been calculated (LANGBEIN and SCHUMM 1958; SCHUMM 1963) using the sediment transport rates of their rivers, or the rates of sediment accumulation in their reservoirs.

An area's yield of sediment depends strongly on its rainfall (LANGBEIN and SCHUMM 1958). Under conditions of effective rainfall of 5″ to 20″/year (most semi-arid regions), sediment yield is high because of the limited amount of sediment and slope-binding vegetation which grows under these dry conditions. With greater rainfalls, the sediment yield decreases because of the increase in vegetation. There is some evidence for an increase in sediment yields at even greater rainfalls (more than 50″/year). Because of the paucity of sediment-binding vegetation in Palaeozoic times (SCHUMM 1968a), we shall assume that the Devonian sediment yield, whatever the precipitation was similar to that of a present-day semi-arid area.

Denudation rates also depend on the size of the drainage area. Large areas tend to have a higher proportion of gentle slopes, and more opportunities for the storage of sediment. Heavy rainstorms are unlikely to effect a whole large area simultaneously. A correction factor for the area of the drainage basin, has been worked out (SCHUMM 1963, p. 3).

Relief within the drainage basin is a third factor which controls denudation rates. SCHUMM (1963, Fig. 2) plotted curves of denudation rate against relief/length ratio of the basins. He drew up 1) a curve representing maximum known denudation rates, and 2) a curve for a number of semi-arid basins underlain by shale and sand bedrock. We suggest that the rates for our Wood Bay Formation might have fallen between these two.

The volume of the Wood Bay Formation provides a direct measure of the amount of denudation which took place, because, as argued above, the basin is known to have been approximately closed. Mean rates for sediment yield of each source area can be worked out by estimating the extent of the alluvium, its thickness, and the duration of Wood Bay times. For some estimates a probable value, and a range of values are given (Table 4 and below, in brackets).

It therefore appears that we can use 0.1 m/1000 years as a general estimate of denudation rate in both areas, and for both climatic models. Using SCHUMM's (1963, Fig. 2) plot of relief/length against denudation rate, we arrive at relief/length ratios of (.01), .03, (.04). This implies the following relief, averaged throughout Wood Bay time:

Western Source areas (50 m), 150 m, (200 m)
Eastern Source areas
(uniform rainfall model) (2000 m), 6000 m, (8000 m)
(contrasting rainfall model) (800 m), 2400 m, (3200 m)

To conclude, relief in the east seems to have been some sixteen to forty times greater than that in the west, on the rainfall pattern assumed. The greater the

Table 4
Estimation of denudation rates

Western areas
Source area 5 by 5 km (25 km²)
Alluvial area (5 by 5 km), 5 by 10 km, (10 by 20 km)
Alluvial thickness 2 km
Therefore, overall erosion (2 km), 4 km, (16 km)
Duration of Wood Bay time, 20 million years
Denudation rate (.1), .2, (.8) m/1000 years
Denudation rate, recalculated (SCHUMM 1963 p. 3) for standard 1500 square mile basin (.05), .1, (.4)

Eastern area (assuming rainfall uniform with western areas)
Source area, 50 by 200 km (10,000 km²)
Alluvial thickness 2.5 km
Alluvial area (150 by 30 km), 180 by 30 km, (200 by 50 km)
Therefore overall erosion (1.1 km), 1.4 km, (2.5 km)
Duration of Wood Bay time, 20 million years
Denudation rate (.06), .07, (.13) m/1000 years
Denudation rate, recalculated (SCHUMM 1963, p. 3) for standard 1500 sq. mile basin (.07), .08, (.15)

Eastern area (assuming rainfall 4 times greater than western area)
Source area, 30 by 80 km (2,500 km²)
Alluvial area (150 by 30 km), 180 by 30 km, (200 by 50 km)
Alluvial thickness 2.5 km
Therefore overall erosion (4.4 km), 5.4 km, (10 km)
Duration of Wood Bay time, 20 million years
Denudation rate (.22), .27, (.5) m/1000 years
Denudation rate, recalculated (SCHUMM 1963, p. 3) for standard 1500 sq. mile basin, (.06), .07, (.13)

relative rainfall in the east, the smaller the source area and relief required to yield the sediment now found.

We shall finally consider (Table 5) the amount of crustal movement which these figures of erosion, sedimentation, and relief imply. We adopt a simplified model for this calculation. The model assumes 1) a single distinct episode of tectonic uplift, followed by 2) erosion, sedimentation and consequent isostatic crustal movement. We assume also that the crust moved by vertical displacement of distinct blocks (rather than tilting or folding), and that the isostatic adjustment consisted of displacement of asthenosphere (density = 3.4), to compensate for movement of lithosphere (density = 2.6).

Our analysis is consistent with initial tectonic uplifts of about 1 km in the west, and 6 km in the east: Erosion, sedimentation and isostatic adjustment finally produced net crustal deformation of amplitudes, 6 km and 9 km respectively.

THE EASTERN MARGIN

Along their eastern margin the Devonian sediments are cut by the Balliolbreen fault, which separates them from the pre-Devonian rocks to the east.

We find no evidence for the existence of a structure along this line in Wood Bay times. In an earlier paper, FRIEND (1967) suggested that it formed a part of a general north-south trending hinge zone contemporaneous with sedimentation. However our further analysis of the sediments, and new work on the subsequent Svalbardian deformation, convince us that no such structure existed along this

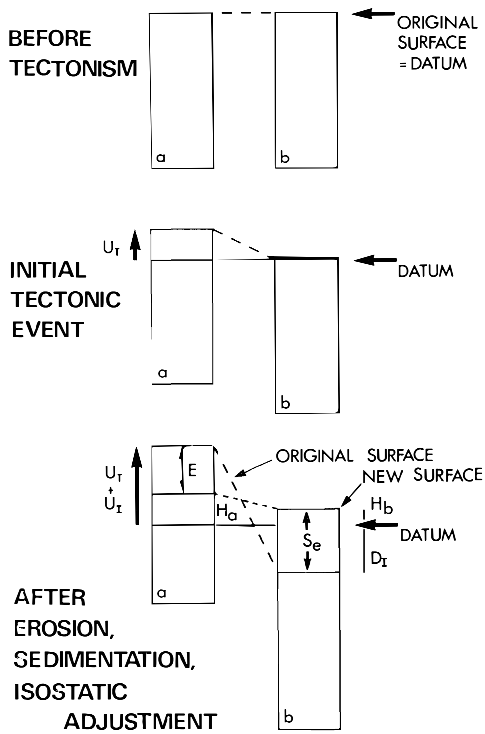

Fig. 34. *Diagram illustrating the model used in estimating relative movement of two crustal blocks (a and b), after an initial tectonic movement of one of them (U_t, tectonic uplift). Erosion (E) on the uplifted block and sedimentation (S_e) on the other block, result in isostatic uplift (U_i) and isostatic downwarp (D_i). After this, the heights of the two blocks, above datum, are H_a and H_b.*

Table 5

Estimation of amount of crustal movement (Fig. 34)

Western area

Erosion (E) of block a, 4 km
Isostatic uplift (U_i), therefore 3.2 km
Sedimentation (Se) on block b, 2 km
Isostatic downwarp (D_i) 1.6 km
Heights (Fig. 34) $H_b = S - D_i = 0.4$ km
$H_a = U_t + U_i - E = U_t - 0.8$ km

1) $H_a > H_b$, or drainage will be reversed
$U_t - 0.8 > 0.4$ km
$U_t > 1.2$ km
2) Final height difference of blocks would be smaller than general relief, which was calculated from denudation rates
$H_a - H_b < 0.15$ km
$U_t < 1.35$ km
3) Initial tectonic uplift must have been greater than general relief
$U_t > 0.15$ km
So $0.15 < 1.2 < U_t > 1.35$ km
Initial tectonic uplift 1.3 km
Total crustal movement $U_t + U_i + D_i = 6$ km

Eastern area (uniform rainfall model)

Erosion (E) of block a, 1.4 km
Isostatic uplift (U_i), therefore 1.1 km
Sedimentation (Se) on block b, 2.5 km
Isostatic downwarp (D_i) 2.0 km
Heights (Fig. 34) $H_b = S - D_i = 0.5$ km
$H_a = U_t + U_i - E = U_t - 0.3$ km

1) $H_a > H_b$, or drainage will be reversed
$U_t > 0.8$ km
2) Final height difference of blocks will be smaller than general relief, which was calculated from denudation rates
$H_a - H_b < 6$ km
$U_t < 6.3$ km
3) Initial tectonic uplift must have been greater than general relief
$U_t > 6$ km
So $0.8 < 6 < U_t > 6.3$, say $U_t = 6$ km
Initial tectonic uplift, therefore, = 6 km
Total crustal movement $= U_t + U_i + D_i = 6 + 1.1 + 2.0 = 9$ km

Eastern area (contrasting rainfall model)

Erosion (E) of block a, 5.4 km
Isostatic uplift (Ui), therefore, 4.2 km
Sedimentation (Se) on block b, 2.5 km
Isostatic downwarp (Di), 2.0 km
Heights (Fig. 34) $Hb = Se - Di = 0.5$ km
$Ha = Ut + Ui - E = Ut - 1.2$ km

1) $Ha > Hb$, or drainage will be reversed
$Ut > 1.7$ km
2) Final height difference of blocks will be smaller than general relief which was calculated from denudation rates
$Ha - Hb < 2.4$ km
$Ut < 3.6$ km
3) Initial tectonic uplift must have been greater than general relief
$Ut > 2.4$ km
So $1.7 < 2.4 < Ut > 3.6$, say $Ut = 3$ km
Initial tectonic uplift, therefore, = 3 km
Total crustal movement $= Ut + Ui + Di = 3 + 4.2 + 2.0 = 9$ km

line during Wood Bay sedimentation. Indeed, as shown above, the local alluvial palaeoslope was oblique to the Balliolbreen fault, with a general dip towards the NNW. We have suggested that the amplitude of the crustal movement, which produced this sedimentation, was as much as 9 km, and this figure stays approximately the same for both rainfall models.

The Svalbardian faulting and associated folding of the Wood Bay Formation occurred at some time during the Upper Devonian (FRIEND 1961; CUTBILL and CHALLINOR 1965). The folding is restricted to a number of north-south trending zones, one just west of the Balliolbreen fault, and two others further west. These zones appear to be the near-surface results of movements in the deeper basement. Their north-south trend probably resulted from the activation of north-south trending structures in the basement.

This Svalbardian deformation provides evidence for an oblique stress similar to that suggested for earlier, Wood Bay times. Within the zones mentioned above, local folds show a tendency to trend NNE-SSW, suggesting compression perpendicular to that direction locally, in a direction which would be compatible with left-lateral strike-slip movement of the zones as a whole. The Balliolbreen fault formed as a slightly later event in this same strain sequence, truncating the local folds. The idea that this was a period of underlying left-lateral strike-slip movement was proposed by HARLAND (1965b, 1969) on grounds of comparison of the general stratigraphies of Spitsbergen, Greenland and Arctic Canada.

To sum up, we regard the Wood Bay and Svalbardian movements as phases of crustal deformation, which both resulted from stresses oblique to the north-south pre-Devonian trend. Firstly, major uplift and downwarp occurred, in response to NNW–SSE compression. Then this was followed by Svalbardian strike-slip fracturing and shearing in response to a similar compression, but controlled by the basement "grain". This is the sort of pattern characteristic of the compressive movement of lithosphere plates where their mutual margin was oblique to the movement. This situation has been described by HARLAND (1971), and called by him "transpression".

In our effort to reconstruct the situation in Wood Bay times (Fig. 35), we have brought the areas of present-day pre-Devonian outcrop in Ny Friesland (HARLAND 1961) and Nordaustlandet (FLOOD and others 1969) further south into a pre-Svalbardian position. The pronounced submarine shelf to the north-west of Nordaustlandet slopes parallel to the Wood Bay palaeoslope. We have, therefore, based our reconstruction on the coincidence of these features, but the arguments for this are not strong. Our study of the sandstones of the Wood Bay Formation provides us with no positive information about the contemporary position of Ny Friesland and Nordaustlandet.

THE WESTERN MARGIN

In the north, the western margin of the Wood Bay Formation outcrop is formed by the Breibogen fault, and there are a number of parallel faults to the south. All these boundary faults are normal, down-throwing to the east. South of the ice-cover of the Holtedahlfonna, a fault which continues this line downthrows to the west. It separates "Lower" and "Middle" Wood Bay outcrops to the east of Orsa-

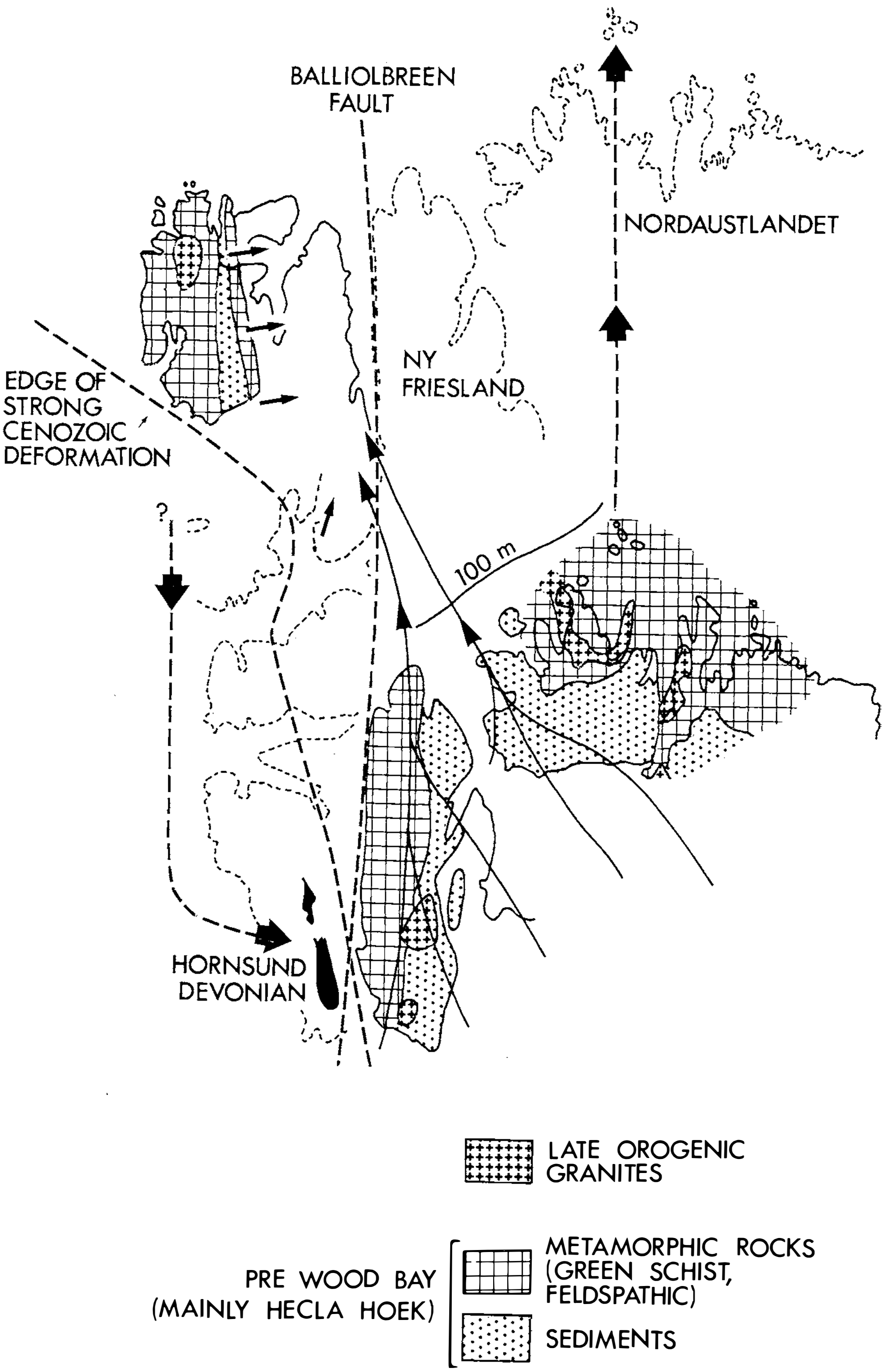

Fig. 35. *Possible palinspastic map of Spitsbergen and Nordaustlandet in Lower Devonian times, before Upper Devonian left-lateral movement of the Balliolbreen fault, and before movement of the Hornsund Devonian in a similar sense. This later movement is indicated by dashed lines and broad arrows. The small arrows show river directions.*

breen from "Upper" outcrops to the west of Orsabreen. These "Upper" outcrops extend westwards as far as Pretender, where there are exposures of an unconformable contact with pre-Devonian, Hecla Hoek rocks. This provides positive evidence of overlap on the western margin of the Wood Bay basin.

Other evidence of the position of the western margin is less direct. In "Lower" times, the palaeocurrent pattern and the pattern of sand members (Figs. 20, 21) show that a western margin parallel to the boundary faults existed as far south as the Holtedahlfonna. Palaeocurrents further south show that the margin there was slightly further west.

In "Middle" rocks there is again palaeocurrent evidence for a margin west of, and parallel to, the present Breibogen fault line.

In "Upper" rocks, this evidence is quite lacking to the north. South of Holtedahlfonna, in the area of "Upper" overlap, palaeocurrents flowed northwards, providing no evidence of a western margin, although 30 km further north deflection of palaeocurrents may indicate some source in that direction.

These points are summarised on Fig. 36, with "minimum" and "possible" areas of "Upper" sedimentation suggested.

Palaeocurrent and overlap information combine to demonstrate the existence of an uplifting source region in north-west Spitsbergen in Wood Bay times. We have considered above the amount of the uplift, and arrived at a tentative figure of 6 km, for total deformation of the crust. This overall figure of 6 km includes a component of 4 km of erosion in the source area. Very minor amounts of orthoclase were derived from this region. It appears likely therefore that the feldspathic lowest unit was not stripped of its overlying 4.5 km (Gee and Hjelle 1966) of marble, pelite and quartzite, at least within the source areas of the western system.

The Breibogen fault cuts the Wood Bay Formation, and therefore, formed later than it. There is no evidence for strike-slip along it. We interpret it as a late-stage dip-slip fracture which occurred when the stress differences caused by uplift, erosion and sediment-loading exceeded the elastic limit of the crust (Walcott, 1970).

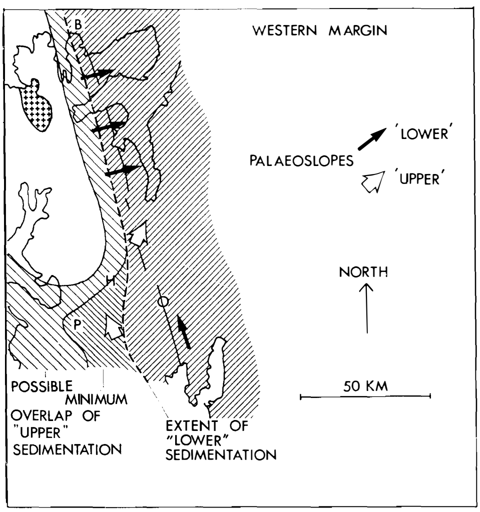

Fig. 36. *Map of western edge of Wood Bay Formation outcrop area showing "Lower" and "Upper" palaeoslope and overlap relations. B, Breibogen; H, Holtedahlfonna; O, Orsabreen; P, Pretender*

THE HORNSUND DEVONIAN

The separate Devonian succession near Hornsund (Fig. 1) has been described by BIRKENMAJER (1964), but has not been included in our main analysis.

Fossil correlation with northern Spitsbergen is limited to 1) thelodont assemblages from Røykensåta, Sørkapp Land, which are very similar to assemblages from the Verdalen Carbonate Member in the north (ØRVIG 1957, p. 288; 1969a, p. 275; 1969b, p. 387). 2) Monaspids from Hornsundtind (BIRKENMAJER 1964, p. 57) which also suggest a correlation with the Stjørdalen Faunal Division of the Wood Bay Formation.

Other than this, correlation has been based on comparison of lithologies. BIRKENMAJER's Marietoppen succession (BIRKENMAJER 1964, p. 49) is 1000 m thick, and consists of a Lower and Middle part (800 m) which he compared with the Lykta and Stjørdalen Faunal Divisions of the Wood Bay Formation of the northern areas, and an Upper part (200 m) which he compared with the Grey Hoek Formation. BIRKENMAJER commented (1964, p. 62) on the lack of sandstone in the succession compared with northern areas. He also reported the lack of scour structures and bed form structures (ripples, cross-bedding, lineation), and he is a geologist experienced in recognising such features. This paucity of both high power structures and bed-load sediment is reminiscent of extreme northern areas, such as Reinsdyrflya. The probable lack of an equivalent of the lower (Kapp Kjeldsen) division of the Wood Bay Formation, and the unconformable contact with the underlying pre-Devonian rocks, are similar to the succession in the overlap area between Pretender and Orsabreen (Fig. 36).

Our work in the north has demonstrated that the Wood Bay Formation there was deposited by eastward and northward flowing river systems. The Hornsund Wood Bay Formation, with its downstream, clay-flat lithologies does not fit into this pattern in its present position. As described above, we follow HARLAND (1969) in believing that there was major left-lateral strike-slip faulting in Spitsbergen in Upper Devonian times. In Hornsund and other south-western areas, there was also major Cenozoic overfolding and thrusting. The lithology and stratigraphy of the Hornsund Devonian provide further evidence of left-lateral strike-slip. An original position for these rocks somewhere to the west of Kongsfjorden (Fig. 35), or even further north, would fit the stratigraphic and environmental trends now recognised in northern Spitsbergen, and require a subsequent strike-slip movement of at least 200 km.

Assessment of methods

This project has involved extending the reconnaissance survey of a large area by carrying out a programme of detailed sedimentological analysis. In this section we ask ourselves whether it was worth it. We firstly review the sorts of information we have gathered, and the assumptions made. We then go on to assess the degree to which our new and detailed knowledge of local Spitsbergen sedimentation contributes 1) to the study of sedimentary processes generally, and 2) to knowledge of the late orogenic tectonic evolution of the area.

SUMMARY OF SORTS OF INFORMATION GATHERED AND DEDUCTIONS MADE

Fieldwork

Elucidation of structure and stratigraphy

Measurement of detailed sedimentological sections, collecting of sandstone samples

First-stage analysis

Palaeocurrent patterns

Sandstone composition patterns

Grain-size variation patterns

Estimation of alluvial variables (Fig. 37)

Channel cross-section and plan. This is based on interpretation of water depth and channel type using sediment features.

Strength of river flow. This is based on use of sedimentary structures and grain-size.

Alluvial slope. This is based on the estimates of the above variables.

Definition and analysis of alluvial systems

Definition of alluvial systems. These are based on palaeocurrents, alluvial variables and sandstone composition.

This leads to generalisation about patterns of water discharge, which results in estimation of areas of drainage basins. This estimation depends on making assumptions about the climate. It also leads to generalisation about patterns of sediment discharge, which results in conclusions about probable source area bed-rock differences.

Analysis of crustal movement

Shape and size of areas of uplift and downwarp. This is based mainly on palaeocurrents, but also on stratigraphical overlap information and water discharge estimates.

Amount of downwarp and uplift. This is based on stratigraphical thickness measurements, calculation of denudation rates (based on area estimates and climatic assumptions), and the assumption of a simplified model of isostatic adjustment.

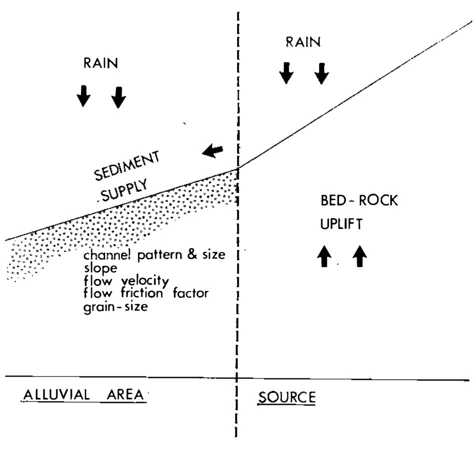

Fig. 37. *Summary of various factors which appear to determine the dependent alluvial variables (lower-case lettering).*

CONTRIBUTIONS TO UNDERSTANDING OF SEDIMENTARY PROCESSES

Our work on the nature of small-scale sedimentary processes, e.g. formation and movement of ripples, depends almost completely on studies in the laboratory or present-day sedimentary environments. This seems inevitable. However, because of the processes of preservation and exposure, ancient sequences often present features which have then usefully become the subjects of further study (e.g. scour structures).

Larger-scale sedimentary processes, e.g. the accumulation of whole alluvial systems, may be studied to great advantage in ancient rock sequences. These sequences may often present information about lateral and vertical variation which would be difficult or impossible to collect from a present-day environment. Specifically in this study, we have been able to gather information about the sedimentation of bed-load and mixed-or-suspended load river channels, clay-flat playas and carbonate lakes, and the interplay of all these environments.

CONTRIBUTIONS TO KNOWLEDGE OF LOCAL TECTONIC EVENTS

One of us (FRIEND 1967) has pointed to the tectonic significance of some of this work in Spitsbergen. The analysis presented in this paper has resulted in some modification of the generalisations made at that stage. We can also now comment, with more experience, on the further potential of this approach.

Our palaeocurrent work is of major importance in defining the positions of uplifting source areas relative to the downwarping alluvial basin. We have in addition, attempted to estimate the size of the uplifting source area. This depends critically on estimation of climatic (particularly rainfall) variation across the area.

We have also been able to estimate the amount of crustal movement, using amongst other features, the pattern of source and alluvial areas already worked out, and work on present-day denudation rates. Both these features depend critically on assumptions about the climate.

We conclude that tectonic information most directly comes from stratigraphical mapping (thicknesses and overlaps), coupled with palaeocurrent work. Other detailed sedimentary work has tectonic value, but this is limited unless coupled with some exploration of the climate factor.

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Appendix 1

Localities and sandstone composition

1) Locality number
2) Grid reference, Northerly
3) Grid reference, Easterly
4) Place name
5) Lower, Middle or Upper Wood Bay Formation
6) Sandstone Composition. *Pairs* of numbers for each sandstone sample

First number orthoclase as percentage of quartz plus all feldspar
Second number (in brackets) is grainsize expressed as divisions of microscope graticule (one division is approximately 0.015 mm). The number is the mean of the longest diameter of the first ten grains in a traverse.

(1)	(2)	(3)	(4)	(5)	6a, (6b)
E 113	340,000	88,250	Georgbreen	U.	0.0(0.5), 0.0(10), 0.0(8.0), 0.0(3), 0.0(2)
E 131	342,000	95,500	Vatnedalen	U.	0.0(0.5), 0.0(0.7)
E 651	241,400	111,600	Oxåsdalen	U.	18.3 (4)
F 411	263,500	125,800	Bulmanfjellet	U.	0.5 (3)
F 413	269,600	125,700	Zeipeldalen	L.	23 (33)
F 431.2	291,100	116,400	Kapp Petermann	L.	—
F 431	282,600	121,900	Simledalen	L.	—
F 445	270,700	114,300	Høegdalen	U.	4.5(4)
F 634	294,700	112,600	Krosspyntdalen	U.	—
F 639.2	291,500	105,200	Landingsdalen	L.	0.0(12)
F 640	290,300	107,400	Landingsdalen	L.	0.2(11)
F 713.1	233,800	127,200	Mimerdalen	L.	32.4(22)
F 942.2	330,000	81,150	Verdalen	U.	—
F 943	327,500	80,500	Verdalen	U.	0.2(10), 0(2.5), 0(2.8), 0(3.5)
F 944	334,000	78,200	Kapp Auguste Viktoria	U.	0.3(2)
F 945	333,900	78,300	Kapp Auguste Viktoria	U.	0.0(3), 0.0(1)
F 1049	225,550	128,000	Billefjorden	M.	3.7(5)
F 1055	223,700	127,700	Billefjorden	M.	1.5(12), 2.8(18)
F 1056	223,400	126,750	Asvindalen	M.	17.9(13)
F 1070	229,900	128,700	Billefjorden	M.	2.4(11)
F 1074	236,000	128,700	Mimerdalen	M.	0.0(6.5), 0.0(6.0)
F 1076	236,400	127,900	Mimerdalen	L.	18.1(5)
F 1078	235,700	127,900	Mimerdalen	L.	25.7(17.6)
F 1080	233,600	127,300	Mimerdalen	L.	38.4(43), 28.4(16)
F 1108	254,800	115,300	Nathorstdalen	L.	20.1(22)
F 1711	336,700	79,000	Prismefjellet	U.	1.1(1.5)
F 1714	334,650	78,900	Prismefjellet	U.	0.0(2)
F 1806	319,000	92,550	Forkdalen	U.	0.0(3.4), 0(4.5)
F 1825	302,300	110,300	Heintztinden	U.	0.0(1.5)
F 1828	296,300	113,900	Krosspyntdalen	U.	1.3(6.3)
F 1847	264,100	125,800	Bulmanfjellet	L.	20.9(12.0)
F 1852	265,500	126,100	Bulmanfjellet	L.	22.1(34)
F 1861	296,550	115,550	Krosspynthytta	M.	2.4(3.1), 0.4(3)
F 2305	249,600	111,300	Lykta	U.	6.7(2), 13(2.5)
F 2345	262,800	109,500	Grønhorgdalen	M.	—
F 2364	280,300	108,900	Skuggefjellet	U.	—
F 2366	275,200	110,100	Hodshalsen	L.	—
F 2390	233,500	125,300	Mimerdalen	L.	3.5(29)
F 2394	233,600	125,900	Mimerdalen	L.	8.3(21)
F 2404.2	241,000	125,300	Reuterskiöldfjellet	L.	7.2(9.5), 2.0(11)
F 2602	242,400	111,700	Tåkefjellet	U.	16.4(4), 15.7(4)
F 2605	247,000	114,500	Triungen	U.	8.2(9.5), 7.8(9.0)
F 2610	245,700	115,700	Triungen	U.	—
F 2612.6	245,600	115,800	Triungen	U.	—
F 2620	250,000	115,000	Citadellet	M.	22.4(10.0), 32.7(13.0), 11.4(5.0), 13.0(16.0)

(1)	(2)	(3)	(4)	(5)	6a, (6b)
F 2623	243,800	117,250	Hugindalen	U.	—
F 2626	245,400	121,800	Stensiöbreen	U.	—
F 2627	251,650	115,700	Citadellet	M.	24.6(14.0), 28.8(48.0), 26.2(34.0)
F 2628.1	251,700	115,600	Citadellet	M.	—
F 2628.2	241,500	97,100	Kapitol	U.	18.7(10.0), 6.9(1.5), 13.1 (10.5)
F 2629	253,150	91,800	Rättvikfjellet	L.	—
F 2632	259,500	90,150	Zornfjellet	L.	0.2(9.5), 0.4(14.0), 0.0(1.5), 0.0(32.0)
F 2634	259,400	89,800	Zornfjellet	L.	0.0(12.0)
F 2637	260,200	89,600	Zornfjellet	L.	—
F 2639	261,600	89,900	Zornfjellet	L.	—
F 2642	258,500	86,400	Siljanfjellet	U.	6.7(13.0), 10.8(4.0)
F 2644	263,800	89,000	Orsafjellet	L.	0.0(8.0), 0.0(28.0), 0.2(8.0)
F 2647	264,800	89,300	Orsafjellet	M.	8.9(5.0)
F 2651	262,000	86,050	Siljanfjellet	U.	—
F 2653	269,200	87,350	Meråkerfjellet	L.	0.0(20.0)
F 2658	266,400	88,300	Orsafjellet	L.	—
F 2659.1	260,900	89,700	Zornfjellet	L.	0.0(28.0)
F 2659.2	257,350	90,000	Rättvikfjellet	L.	0.2(7.0)
F 2661	254,400	79,350	Falunfjellet	U.	14.2(7.0)
F 2663	260,700	74,000	Vortefjellet	U.	—
F 2664	255,600	75,800	Palasset	U.	8.8(4.0), 20.4(9.0)
F 2665	260,000	81,900	Särnafjellet	U.	3.2(2.5)
F 2667	243,000	96,700	Garborgnuten	U.	15.9(5.5), 7.6(9.0), 26.5(10.0), 16.4(5.5)
F 2668	240,100	96,700	Kapitol	U.	17.1(5.0)
F 2669	247,100	96,100	Gygrestolen	U.	9.3(6.0), 8.6(4.5), 11.4(7.0)
F 2670	251,600	104,900	Raudkollen	U.	15.3(10.0), 20.6(8.5), 15.3(6.5)
F 2671	269,400	104,600	Stjernspetzfjellet	M.	13.4(8.0), 2.9(8.0), 11.8(6.0)
F 2672	270,200	99,400	Venfjellet	M.	17.2(9.0), 9.0(4.5), 14.3(12.0)
F 2674	274,800	96,400	Stauptindane	M.	7.0(4.0)
F 2675	262,250	111,900	Grønhorgdalen	L.	13.8(18.0), 13.2(3.0), 0.2(18.0)
F 2676	262,000	111,800	Grønhorgdalen	L.	18.0(19.0)
F 2677	263,800	122,500	Bulmanfjellet	L.	29.2(14.0), 23.2(55.0), 28.1(11.0)
F 2678	266,000	124,800	Bulmanfjellet	L.	22.9(11.0), 25.1(23.0), 25.1(24.0)
F 2680.4	262,100	127,000	Odellfjellet	U.	22.6(9.0)
F 2681	262,200	124,800	Bulmanfjellet	L.	32.6(15.0), 26.8(13.0)
F 2682	299,850	112,150	Heintztinden	U.	0.2(2.5), 0.2(9.5), 0.0(2.0)
F 2684	293,500	113,200	Errol Whitefjellet	U.	0.8(2.0), 5.7(5.5)
F 2685	283,000	113,600	Ove Dahlfjellet	U.	19.8(10.5), 7.4(30.0)
F 2686	276,100	111,400	Sir Thomasfjellet	L.	15.3(13.0), 0.0(28.0), 7.4(25.0), 10.0(10.5)
F 2686.2	277,850	118,100	Dyrskardet	L.	30.8(53.0), 24.5(45.0)
F 2687	283,300	121,800	Simledalen	L.	17.1(22.0)
F 2689	278,150	122,500	Ridderborgen	L.	23.9(14.0), 34.3(12.0), 23.3(12.0)
F 2690	283,600	121,300	Gråkammen	L.	16.5(6.0), 8.5(10.0), 13.9(4.5)
F 2691	277,800	114,300	Sir Thomasfjellet	M.	12.2(22.0), 3.7(7.0), 1.2(5.0), 5.2(9.0)
F 2692	272,100	107,500	Blindernbreen	M.	12.5(11.0)
F 2693	256,900	110,000	Rebbingen	U.	18.9(12.0), 8.8(8.5)
F 2695	249,200	110,700	Lykta	U.	10.8(4.5), 9.9(5.0)
G 201	268,800	124,700	Purpurfjellet	M.	14.8(2.0)
G 204	286,700	117,100	Jørgensendalen	M.	0.2(4.5), 0.2(3.5), 0.7(6.5)
I 2	242,700	117,200	Hugindalen	U.	13.3(9.0)
I 4	242,150	118,100	Hugindalen	U.	5.3(3.0)
K 308	327,600	103,500	Forkdalen	U.	0.0(1.0)
K 309	326,900	103,400	Forkdalen	U.	0.0(1.5), 0.0(0.5)
M 1207	327,700	81,400	Verdalen	U.	—
M 1208	333,400	77,800	Kapp Auguste Viktoria	U.	—
M 1502	360,700	59,570	Skjoldkollen	M.	4.8(4.0)
M 1508	360,550	59,690	Skjoldkollen	M.	9.8(4.0), 5.7(4.5)
M 1513	360,870	59,770	Skjoldkollen	M.	11.6(2.5), 10.1(1.7), 3.6(7.0)
M 1516	325,750	80,150	Sørlifjellet	U.	0.3(3.0)
M 1571.2	354,200	58,200	Sørkollen	M.	—
M 1572	353,280	57,580	Løyningdalen	M.	0.0(8.0)
M 1573	350,500	58,050	Fotkollen	M.	—
M 1575	357,300	53,600	Arlahaugen	L.	0.7(70.0)
M 1578	361,470	61,900	Skjoldkollen	M.	—

(1)	(2)	(3)	(4)	(5)	6a, (6b)
M 1582	356,890	60,090	Nordkollen	U.	—
M 1587	358,250	59,520	Nordkollen	U.	—
M 1592	360,670	62,010	Skjoldkollen	U.	0.0(1.5)
M 1595	358,810	59,910	Skjoldkollen	U.	—
M 2007	350,000	56,070	Fotkollen	L.	0.0(35.0), 0.2(35.0), 0.0(65.0), 7.1(40.0)
M 2008	348,030	56,610	Siktefjellet	L.	2.1(10.0)
M 2046	336,620	61,200	Keisar Wilhelmhøgda	L.	0.2(11.0), 0.6(9.0), 0.0(10.0)
M 2049	339,650	65,030	Roosfjella	L.	0.2(2.0)
M 2052	340,320	61,400	Liefdefjorden	L.	0.0(35.0), 0.0(12.0), 1.6(13.0)
M 2054	340,500	64,630	Roosfjella	L.	3.0(1.8)
M 2056	335,100	62,710	Roosfjella	L.	0.4(23.0), 0.4(20.0)
M 2062	330,000	64,670	Bockfjorden	L.	—
M 2064	330,090	65,150	Sjøværnbukta	L.	0.0(12.0), 0.0(3.5), 0.3(18.0)
M 2068	315,220	69,670	Karlsbreen	L.	—
M 2069	327,000	69,000	Næssöpynten	L.	—
M 2070	323,000	69,000	Vulkanhamna	L.	—
M 2071	323,350	71,160	Kronprinshøgda	M.	0.2(1.5)
M 2072	321,960	70,810	Kronprinshøgda	M.	—
M 2073.1	322,200	69,200	Vulkanhamna	L.	—
M 2073.2	322,400	73,100	Halvdandalen	L.	—
M 2073.3	324,200	72,500	Halvdandalen	L.	—
M 2074	303,820	71,400	Sigurdfjellet	L.	—
M 2075	355,500	77,000	E. Reinsdyrflya	U.	0.0(7.0)
M 2100.1	350,170	73,750	Stasjonsøyane	M.	—
M 2100.2	347,450	68,560	Andøyane	M.	—
M 2100.3	341,860	69,450	Måkeøyane	L.	—
M 2101.1	347,500	68,800	Andøyane	M.	0.0(3.0)
M 2105	321,690	68,650	Karlsbreen	L.	—
M 2106	323,150	69,090	Bockfjorden	L.	—
M 2108	318,230	69,290	Karlsbreen	L.	—
M 2109	319,630	69,200	Karlsbreen	L.	—
M 2110	319,800	69,920	Karlsbreen	L.	—
M 2111	317,050	69,580	Karlsbreen	L.	0.0(18.0), 0.0(15.0), 0.9(4.0)
M 2112	317,300	69,610	Karlsbreen	L.	0.0(11.0), 2.9(1.5)
M 2113	317,160	69,650	Karlsbreen	L.	—
M 2114	316,760	69,610	Karlsbreen	L.	5.8(2.0)
M 2115	321,910	68,790	Karlsbreen	L.	—
M 2116	321,420	68,760	Karlsbreen	L.	—
M 2122	310,450	75,450	Risefjella	L.	0.0(7.5)
M 2123.1	311,360	82,840	Scott Keltiefjellet	M.	2.7(1.5)
M 2123.2	311,360	82,840	Scott Keltiefjellet	U.	0.7(3.0)
M 2124	311,360	82,540	Scott Keltiefjellet	M.	0.2(3.5), 1.4(1.0), 2.0(1.5), 2.8(2.0)
M 2125	307,460	82,450	Vaktaren	U.	16.1(9.0), 6.4(2.0)
M 2127	302,500	83,100	Woodfjorddalen	M.	—
M 2128	307,880	76,330	Sigurdfjellet	L.	0.8(4.5)
M 2131	296,800	82,000	Wagnerfjellet	M.	—
M 2132	297,100	81,590	Wagnerfjellet	M.	—
M 2134	289,650	79,320	Sverresborg	U.	11.0(1.5), 19.6(6.0), 19.8(3.5)
M 2135	297,230	79,870	Grevefjellet	M.	7.8(1.0), 0.8(4.5)
M 2137	300,020	79,670	Grevefjellet	L.	0.3(1.5)
M 2138	296,610	74,450	Svartpiggen	L.	0.0(20.0), 0.8(15.0), 1.8(4.5)
M 2139	299,950	88,270	Nidhogg	U.	1.1(4.0), 2.3(9.0)
M 2141	296,630	86,050	Marinova	M.	2.3(1.5)
M 2141.2	294,700	94,700	Kulissene	U.	—
M 2143	301,460	77,410	Grevefjellet	M.	—
J 537	331,790	65,750	Sjøværnbukta	L.	—
U 017.2	291,100	116,400	Kapp Petermann	L.	0.2(9.0)
Y 107	286,000	117,700	Jørgensendalen	M.	0.2(2.0)
Y 108	275,300	121,000	Høegdalen	L.	—
Y 109.2	267,300	123,000	Zeipeldalen	M.	—
Y 519	360,966	54,158	Arlahaugen	M.	0.6(14.0)
Y 521	358,940	58,870	Nordkollen	M.	—
Y 526	359,300	53,100	Arlahaugen	L.	—
Y 540	354,120	57,990	Sørkollen	M.	0.5(2.0), 5.2(1.5)

Appendix 2

Palaeocurrent data

1) Locality number – see Appendix 1 for details
2) Number of observations
3) Probability that distribution is random is smaller than this figure (see Curray, 1956, Fig. 4), *not significant at 95% level of confidence
4) Mean azimuth (angle to True North in degrees)
5) 95% confidence limits on mean azimuth in degrees

Lower (31 localities)

(1)	(2)	(3)	(4)	(5)
F 2629	35	10^{-3}	332	26
F 2632	43	10^{-5}	16	19
F 2644	73	10^{-15}	356	12
F 2653	23	5×10^{-2}	359	36
F 2659.1	9	10^{-3}	12	15
F 2659.2	3	5×10^{-2}	350	20
F 2675	13	10^{-2}	327	28
F 2676	9	5×10^{-2}	323	45
F 2677	93	10^{-5}	333	15
F 2678	44	10^{-5}	320	20
F 2686	84	10^{-10}	357	11
F 2689	85	10^{-20}	326	10
F 2690	62	10^{-10}	308	12
M 2007	10	10^{-2}	44	11
M 2049	5	2×10^{-2}	14	17
M 2052	8	10^{-3}	43	24
M 2056	15	10^{-4}	43	22
M 2062	22	10^{-4}	71	12
M 2068	10	2×10^{-2}	34	36
M 2073.2	9	2×10^{-1}*	38	57
M 2074	8	10^{-2}	75	35
M 2100.3	18	10^{-3}	26	25
M 2105	19	10^{-3}	28	19
M 2106	13	10^{-4}	57	9
M 2109	7	10^{-2}	28	26
M 2111	24	10^{-5}	45	13
M 2122	12	10^{-3}	21	24
M 2128	11	10^{-4}	31	7
M 2138	8	10^{-2}	57	29
J 537	10	10^{-2}	29	36
Y 526	6	10^{-2}	55	25

Middle (21 localities)

(1)	(2)	(3)	(4)	(5)
F 1861	55	10^{-10}	329	14
F 2344.2	15	10^{-4}	353	23
F 2620	18	10^{-4}	342	20
F 2627	11	10^{-3}	294	19
F 2671	84	10^{-5}	320	14
F 2672	64	10^{-5}	336	18
F 2674	37	10^{-5}	314	17
F 2691	46	10^{-5}	315	15
F 2692	27	10^{-4}	342	21
M 1502	10	5×10^{-2}	40	47
M 1513	19	10^{-3}	5	25

Middle (21 localities continued)

(1)	(2)	(3)	(4)	(5)
M 1578	8	5×10^{-2}	324	54
M 2071	8	10^{-2}	15	33
M 2072	15	10^{-3}	30	23
M 2100.1	12	10^{-4}	13	19
M 2100.2	12	5×10^{-2}	221	48
M 2123	31	10^{-5}	5	7
M 2135	12	10^{-4}	13	18
M 2141	12	4×10^{-2}	360	39
M 2143	13	10^{-4}	28	12
Y 540	7	10^{-2}	118	27

Upper (31 localities)

(1)	(2)	(3)	(4)	(5)
F 1116	11	10^{-1}	322	48
F 1829	12	10^{-2}	345	31
F 2305	42	10^{-3}	4	24
F 2602	11	10^{-2}	326	31
F 2605	34	10^{-5}	346	20
F 2610	42	10^{-10}	339	16
F 2623	19	10^{-2}	313	32
F 2626	4	5×10^{-2}	299	45
F 2628.2	32	10^{-4}	344	24
F 2642	18	10^{-4}	353	24
F 2651	17	10^{-5}	340	20
F 2661	43	10^{-5}	348	18
F 2663	33	10^{-5}	335	18
F 2664	60	10^{-10}	314	14
F 2665	52	10^{-5}	311	18
F 2667	64	10^{-10}	7	15
F 2668	21	10^{-4}	7	30
F 2669	81	10^{-5}	327	15
F 2670	68	10^{-10}	330	14
F 2680	8	2×10^{-1}*	329	60
F 2682	87	10^{-5}	332	13
F 2684	62	10^{-5}	384	15
F 2685	23	10^{-2}	326	30
F 2693	22	10^{-5}	337	15
F 2695	25	10^{-3}	353	25
I 2	5	3×10^{-1}*	333	62
M 1587	10	5×10^{-2}	282	44
M 2125	10	2×10^{-2}	23	41
M 2134	11	10^{-2}	40	27
M 2139	16	10^{-3}	359	26
M 2141.2	9	2×10^{-2}	9	38

Appendix 3

Sedimentological sections and grain-size analysis

1) Locality number – see Appendix 1 for details
2) Length of section (m)
3) Mean Grain size (Ø) per 0.25 m. interval of section (= "weighted" mean grain-size)
4) *Semi-cycle,* mean maximum grain-size (Ø)
5) *Semi-cycle,* mean proportion of sand (%) per semi-cycle
6) *Sand semi-cycle,* mean maximum grain-size (Ø)
7) *Sand semi-cycle,* mean thickness of semi-cycle (m)
8) *Sand semi-cycle,* mean thickness of sand member (m)

* no complete sand semi-cycle present (no sand member, or only one sand member).
NC: section not continous enough to provide semi-cycle statistics.

Lower (38 sections)

(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
F 2366	28	1.4	0.8	75	0.7	2.1	1.7
F 2629	53	4.2	3.2	18	2.0	5.3	2.0
F 2634	26	3.7	2.1	42	2.0	2.2	0.6
F 2639.2	63	3.6	1.6	35	1.2	5.6	1.8
F 2653	48	4.4	3.8	6	2.5	14.5	1.1
F 2658	34	4.6	3.4	16	3.0	7.9	0.8
F 2659.1	22	3.0	2.3	53	1.6	2.3	1.2
F 2659.2	148	4.3	2.7	24	1.9	5.2	0.8
F 2675	48	2.4	1.6	80	1.2	3.9	2.9
F 2676	121	3.7	3.2	30	2.3	5.4	2.1
F 2677	235	1.6	1.0	90	1.0	4.2	3.9
F 2678	470	1.6	0.9	93	0.8	3.5	3.4
F 2686	409	3.0	2.9	43	2.0	5.5	3.4
F 2687	18	1.3	0.5	100	1.0	6.0	6.0
F 2689	261	1.5	0.7	96	0.7	4.6	4.4
F 2690	180	2.0	1.8	82	1.6	4.5	4.0
M 2052	36	4.2	4.0	1	3.9	21*	7.0
M 2054	61	4.7	3.6	1	2.7	14.5	0.1
M 2056	160	4.7	3.3	11	3.3	8.7	0.8
M 2064	23	3.4	1.0	22	-0.9	2.5	0.6
M 2069	12	4.6	4.0	0	4.0	14*	0.7
M 2070	30	4.3	1.8	9	1.0	4.4	0.5
M 2073.1	25	4.8	4.0	0	4.0	23*	0.8
M 2073.3	24	4.4	4.0	0	4.0	22*	0
M 2105	43	4.2	-0.2	30	-0.1	3.6	0.9
M 2106	39	4.0	1.5	14	-1.0	4.9	0.7
M 2108	49	4.6	2.9	10	1.7	5.5	0.6
M 2109	35	3.9	1.2	44	0.9	1.8	0.7
M 2110	37	4.6	2.1	10	7.0	3.0	0.4
M 2111	49	3.8	0.3	56	0.21	1.9	0.6
M 2112	47	4.5	2.3	22	-0.3	6.3	0.6
M 2113	57	3.9	0.9	48	-0.3	2.2	0.6
M 2114	62	4.4	1.4	24	-0.1	4.1	0.7
M 2115	59	4.5	1.0	50	0.9	2.7	0.6
M 2116	68	4.2	0.6	53	-0.2	3.2	0.6
M 2128	44	4.7	2.1	5	1.0	9.1	0.3
M 2137	51	4.7	3.3	0	-1.0	25.5	0.1
M 2138	60	4.3	1.5	13	0.3	4.5	0.5

Middle (23 sections)

(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
F 1861	141	3.0	0.8	70	0.8	4.0	2.1
F 2345	49	2.8	2.4	24	0.0	12.4	7.0
F 2628.1	15	3.3	0	59	7.0	9.0	6.0
F 2671	209	3.6	2.7	39	1.9	3.7	1.8
F 2672	273	4.1	3.1	16	1.5	7.3	1.7
F 2674	109	3.9	3.4	24	2.9	3.2	1.1
F 2691	198	3.4	2.8	30	1.5	5.8	2.5
F 2692	88	3.8	3.4	14	1.4	8.7	3.0
M 1502	87	4.6	3.2	1	0.5	27.8	0.8
M 1508	208	4.7	3.3	12	1.2	21.9	2.3
M 1513	103	4.5	3.3	10	0.6	9.7	4.5
M 1571.2	73	4.8	4.0	0	4.0	25*	0
M 1572	29	3.7	2.0	16	3.0	13.1	0.2
M 1573	37	4.6	3.6	6	3.0	8.1	1.0
M 1578	104	4.7	3.5	4	1.0	21.8	0.8
M 2123	74	4.8	4.0	0	4.0	25*	0
M 2124	65	4.5	3.5	4	0.0	10.7	0.6
M 2127	23	4.7	4.0	0	4.0	25*	0.3
M 2131	22	4.7	4.0	0	4.0	22*	0
M 2132	43	4.8	3.9	1	4.0	22*	0.7
M 2135	52	4.8	3.3	2	−7.0	10.8	0.2
Y 519	45	4.8	1.5	38	−7.0	0.4	0.4
Y 521	47	4.7	4.0	0	4.0	6.8	2.3

Upper (33 sections)

F 634	22	3.2	3.3	25	3.0	8.9	2.7
F 942.2	30	4.7	3.0	10	3.0	17.0	2.0
F 943	105	4.1	2.3	24	1.54	7.0	1.3
F 1822	12	4.0	3.3	9	2.0	3.5	0.2
F 2305	81	4.1	2.7	32	1.9	7.6	2.3
F 2364	38	3.7	2.9	14	1.3	10.4	3.7
F 2605	37	3.3	2.9	24	2.0	3.8	0.8
F 2612.6	12	3.3	2.5	48	3.0	2.5	0.5
F 2626	10	4.6	2.8	26	2.7	1.6	0.5
F 2628.2	15	4.5	3.3	17	2.5	5.9	1.1
F 2661	88	4.3	3.7	10	2.8	12.7	1.0
F 2663	105	4.4	3.6	12	2.6	10.9	1.3
F 2664	165	4.4	3.4	12	2.1	10.4	1.2
F 2665	50	4.1	3.8	8	3.0	7.0	0.5
F 2667	89	3.2	2.7	35	1.9	6.1	2.6
F 2668	46	3.7	3.6	16	2.7	6.8	3.1
F 2669	98	2.9	1.2	30	2.0	2.0	0.8
F 2670	156	3.6	3.2	36	2.6	5.2	2.0
F 2680.4	70	1.6	0.2	94	−0.08	2.6	2.6
F 2682	268	4.1	3.2	20	1.9	5.4	1.2
F 2684	288	4.1	3.6	14	2.5	6.5	1.9
F 2685	108	3.2	2.5	44	1.4	4.7	2.5
F 2693	39	2.9	1.9	65	0.7	3.1	2.8
F 2695	50	3.1	NC	NC	NC	NC	NC
I 2	24	4.1	2.8	28	3.0	2.6	0.8
M 1207	45	4.6	3.3	9	2.5	15.2	1.1
M 1208	29	4.9	3.3	4	4.0	21*	0.3
M 1516	97	4.6	2.6	9	2.0	5.3	2.0
M 1582	37	4.9	4.0	0	4.0	12*	1.1
M 1587	39	4.9	3.7	4	4.0	24*	2.0
M 1592	47	4.9	3.5	7	4.0	35*	1.2
M 1595	33	4.8	3.0	15	3.0	4.5	1.0
M 2134	60	4.4	1.9	10	−0.3	6.5	1.7

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