PROCEEDINGS

OF THE

USSHER SOCIETY

VOLUME TWO
PART SIX

Edited by
E. B. SELWOOD

REDRUTH, DECEMBER 1973

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CONFERENCE OF THE USSHER SOCIETY HELD AT WEYMOUTH, JANUARY, 1973

CHAIRMAN'S REPORT

The decision to hold the conference at Weymouth was, in the event, a happy one and provided the Society with an opportunity to examine under the guidance of Professor M. R. House some of the younger rocks in the South-west. In keeping with this venture into new terrain was the invited address from Dr. Raymond Casey of I.G.S., who spoke on 'Dorset from a distance,' and set the tone for the highly informative, enjoyable and expertly presented proceedings that were to follow. To Dr. Casey, as to Professor House, the Society is much indebted for leading it into fresh fields.

The attendance, somewhat under 50, represented many aspects of interest in the geology and evolution of South-west England and this was reflected in the varied subjects of the papers presented. It is difficult to single out special items for mention but the Friday afternoon seminar by Drs. Howarth and Lowenstein on Geochemical Reconnaissance in S.W. England aroused much interest in field and analytical techniques in geochemistry, in computerisation of data and map production and in the implications of the geochemical results.

At the Annual General Meeting there was some discussion about the venue for the 1974 meeting. It was finally agreed that the next conference should be held in Cornwall and that a specialist in the geology of Britanny should be invited from France to address the Society. A field excursion to Britanny might perhaps be arranged for the following summer (1974).

From the Weymouth meeting I gained the firm impression that the Ussher Society is as vigorous as ever. The many facets of geology displayed in South-west England are most ably polished by their investigators. Since its inception in the early 1950's the annual meeting of geologists and geomorphologists working in the

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South West, it seems to me, has provided the right kind of market place for the exchange of geological goods and gossip and the trade is as brisk and sharp as any elsewhere. On a future occasion I would like to say more about the particular contribution our Society has made to promoting geological activity in Britain but in the meantime I look forward to the next meeting and the possibility of "going into Europe."

D. L. Dineley.

PRIMARY REGIONAL GEOCHEMICAL RECONNAISSANCE OF SOUTH-WEST ENGLAND

by P. L. Lowenstein and R. J. Howarth

The Applied Geochemistry Research Group is currently completing a geochemical survey of England and Wales, financed by the Wolfson Foundation, and plans to publish maps for some 20 major and trace elements in 1974. The Atlas should yield information of use to the agricultural and mining industries in addition to geological, biological and related environmental research.

The complete survey has involved the collection of some 50,000 samples of stream sediment collected from tributary drainage at an average density of one sample per square mile. Stream sediments were chosen because they approximate more or less closely to a composite sample of the products of weathering derived from the rocks and soils upstream from the sample site. The samples have been analysed using an A.R.L. 29000B automatic spectrograph for Al, Ba, Ca, Co, Cr, Cu, Fe, Ga, K, Li, Mg, Mn, Ni, Pb, Sc, Si, Sri, Sr and V, and for As, Cd, Mo and Zn by rapid chemical procedures commonly used for prospecting purposes.

Analytical and sampling errors varied considerably according to the element, its concentration and the nature and origin of tile sample. Mean precisions for the spectrographic analyses were of the order of ± 50 per cent at the 95 per cent confidence level, and ±25 per cent for the chemical determinations. Mean sampling error is difficult to estimate but, judging from the results of limited replicate sampling, the ranges of the sampling and spectrographic analytical errors are comparable.

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The data were then computer processed to obtain grey-scale symbol maps on a line-printer using a special- plotting program which incorporates moving-average smoothing. The purpose of the smoothing was to reduce the 'noise' produced by sampling and analytical errors and local small-scale geochemical features. The smoothed data were then processed using an LGP2703 Lasergraphic plotter to produce maps with a true photographic grey-scale.

In South-west England the maps for several of the elements show a good relationship to the geology, both for igneous and sedimentary formations. Many elements show pronounced halos in the area of the metamorphic aureoles surrounding the granites. An outstanding feature is the surprisingly extensive pattern of strontium values centred around the Mendip Hills in the north-east. The exceptionally strong patterns for tin and arsenic, and the less striking anomalies for lead and copper are centred on the mineralised districts of Devon and Cornwall and, in the case of lead, the Mendip area also. The peak values are mostly related to past and present mining and industrial activity but in places extend over a larger area than would be expected from the known distribution of the lode deposits.

Extensive patterns of high molybdenum associated with black shale horizons outline areas in which molybdenum-induced copper deficiency may occur in grazing cattle, and patterns at potentially deficient levels for copper, cobalt and manganese could be a contributing factor to subclinical disorders affecting agricultural productivity.

These maps aim to delineate the broad-scale patterns of element distribution and should be regarded as experimental prototypes which, it is hoped, will not only have immediate practical value but will stimulate additional geochemical research.

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MUSCOVITE (‘GILBERTITE’) FROM THE MELDON APLITE

by M. Nawaz Chaudhry and R. A. Howie

Abstract. The chemistry, and optical and physical properties of two so-called gilbertites from joint faces in the Meldon aplite-pegmatite, near Okehampton, Devon, show that they are similar to normal muscovite.

The Meldon aplite is a soda-lithia rich aplite dyke about 20-25 m thick in the Lower Culm Measures about 1 km north-west of the main Dartmoor granite (Worth 1920). Albite (Chaudhry 1971), quartz, lithium-aluminium micas (Chaudhry & Howie 1973) and K-feldspar (Chaudhry 1971) are the essential minerals, and elbaite, topaz (Chaudhry & Howie 1970), fluorite, apatite and petalite occur as accessories. Rarely a pale yellow-green secondary mica, similar to that from South-west England referred to as ‘gilbertite,’ occurs along joints and fractures but does not occur in any other way in the Meldon aplite. It appears to have been derived by hydrothermal-pneumatolytic replacement. Chemical analyses and optical and physical properties of two samples are given in Table 1.

Although the gilbertites are associated with a Li-bearing suite of minerals they appear to be Li-free (below the level of detection on the flame-photometer). XRF analyses, however, show that they are relatively rich in rubidium and have a high Rb/Cs ratio. Like ordinary muscovites they are rich in water and relatively poor in fluorine. The yellowish green colour may be attributable to their appreciable (0.22 and 0.28%) content of FeO whereas Fe_2O_3 is absent. Their refractive indices although lower than average fall within the muscovite range: their *b* cell dimensions are somewhat lower than normal values. Their optic axial angles are large and close to the upper limit for muscovite.

Gilbertite was named by Thomson (1831) in honour of Mr. Davies Gilbert, late President of the Royal Society, the name being applied to a mineral previously called 'Cornish talc,' collected by Thomson from near St. Austell. There seems no reason to retain the name.

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TABLE 1. Chemical analysis of gilbertitic muscovite from the Meldon aplite, near Okehampon, Devonshire.

			Ions per 24 (O,OH,F)				
SiO_2	44.23	44.67	Si	5.934	8.00	5.953	8.00
TiO_2	0.14	0.12	Al	2.066		2.047	
Al_2O_3	39.75	38.64					
Fe_2O_3	0.00	0.00	Al	4.221	4.31	4.024	4.13
FeO	0.22	0.28	Ti	0.015		0.012	
MnO	0.27	0.35	Fe^{3+}	0.000		0.000	
MgO	0.08	0.12	Fe^{2+}	0.025		0.031	
CaO	0.12	0.45	Mn	0.031		0.039	
Li_2O	0.00	0.00	Mg	0.016		0.024	
Na_2O	0.58	0.52	Li	0.000		0.000	
K_2O	9.74	9.80					
Rb_2O	0.60	0.58	Ca	0.017	1.89	0.064	1.91
Cs_2O	0.03	0.00	Na	0.152		0.135	
F	1.15	0.85	K	1.667		1.666	
H_2O^+	3.14	3.83	Rb	0.052		0.500	
H_2O^-	0.29	0.34	Cs	0.002		0.000	
	100.34	100.55					
F = O	0.53	0.37	OH	2.810	3.30	3.405	3.76
Total	99.81	100.18	F	0.488		0.358	
β	1.585	1.584					
2Vα	48°	47°					
Sp.gr.	2.84	2.83					
Polytype	$2M_1$	$2M_1$					
b(Å)	8.98	8.99					
c sin β (Å)	20.11	20.11					

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Geochemical and preliminary palaeomagnetic results of the Lemail lamprophyre, Wadebridge, Cornwall (Abstract) :

by M. E. Cosgrove and N. Hamilton.

It has long been suggested that the lamprophyre dykes of S.W. England are genetically related to the Permian volcanics of eastern Devon, perhaps occupying feeder channels to a once more extensive volcanic field. The results of geochemical investigations of some of these lamprophyres supports such a relationship.

The Lemail lamprophyre, very well exposed in a railway cutting, was selected for detailed study. It is a typical minette with well developed biotites and orthoclase and with a chemistry very similar to that of the Tiverton group of minettes and lamproites of the Permian volcanics. In particular it shows richness in TiO_2, MgO and K_2O, with a trace element suite characterised by high P, Ni, Rb, Zr, Nb, Ba, R.E's., and Th values.

A preliminary result for the direction of magnetization of the Lemail dyke (50° 31' 22" N, 4° 47' 22' W) has been obtained from measurement of the natural remanent magnetization (N.R.M.) of 18 cylindrical specimens cut from two oriented blocks. The measurements were made on a low frequency complete results spinner magnetometer. The remanent intensities are in the range 0.7 - 4.0 x 10' Gauss/cm^3. A mean direction of magnetization, excluding one unreliable specimen, is given by Declination 179.3°, Inclination 58.1°, K = 9.3, $\alpha_{9.5}$ = 12.3°, with respect to present-day horizontal and true north.

This direction, particularly the inclination, is widely different from that of the possible contemporaneous Exeter volcanics as reported by Zijderveld (1967). The discrepancy may be related to an unstable component or more likely to the mode of intrusion or subsequent tilting of the dyke. Further studies are in progress including magnetic cleaning.

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DISTRIBUTION OF F AND Cl IN SOME CONTACT AND REGIONALLY METAMORPHOSED CORNISH GREENSTONES

by P. A. Floyd and R. Fuge

Abstract. A comparison of low-grade regional metamorphosed greenstones and contact metamorphosed hornfelses from Cornwall show a marked increase in F and Cl in the latter group. With increasing grade of contact metamorphism (albite-epidote hornfels to hornblende hornfels facies), F shows a significant increase from 743 to 1087 p.p.m., while Cl only exhibits a small increase from 1207 to 1355 p.p.m. The regionally metamorphosed greenstones have higher Cl and lower F values compared with other unmetamorphosed basic rocks. Two main trends are recognised in the F-Cl relationship for basic rocks : (a) antipathetic, for un metamorphosed and slightly metamorphosed basic rocks (high Cl-low F sample relationship), and (b) sympathetic, for higher grade contact metamorphosed basic rocks (high Cl-high F sample relationship).

1. Introduction

This report presents the preliminary results of a survey of the distribution of F and Cl in metamorphosed basic rocks from Cornwall. In view of the F and Cl rich nature of the Cornubian granite batholith (Fuge and Power 1969), contact metamorphosed basic hornfelses adjacent to the Land's End granite were sampled so that a comparison could be made with regional greenstones elsewhere. Also, some metasomatically altered hornfelses, with anthophyllite, cummingtonite and calc-silicates (Floyd 1965) were analysed to see if any variation in the distribution of F and Cl took place during metasomatism.

The average results for the various contact hornfelses compared with the regional Cudden Point greenstone are shown in Table 1.

2. Contact hornfelses

As expected both the low grade actinolite hornfelses (albite-epidote hornfels facies) and the higher grade hornblende hornfelses (hornblende hornfels facies) have much higher average values for F and Cl than the low-grade regional dolerite-diabase sheet from

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Cudden Point. Clearly a much greater proportion of Cl has been added than F. In general, the higher the hydrous mineral content (amphibole and biotite) of the hornfelses, the higher the F; suggesting that this element is replacing the hydroxyl group in these minerals. Cl, however, seems more erratic in behaviour, although in the regionally metamorphosed greenstones there is a tendency for the highly altered chlorite-rich dolerites to have generally lower values.

With progressive contact metamorphism a definite increase in F is seen, whereas only a minimal increase in Cl content is observed.

The ferromagnesian amphibole-bearing metasomatic hornfelses show similar F values to their metamorphic parents-the hornblende hornfelses. The higher value for the calc-silicate metasomatic hornfelses is probably due to the presence of minor fluorite. Cl on the other hand, shows a considerable and progressive decrease from the parental hornfelses to the cummingtonite-bearing and anthophyllite-bearing hornfelses respectively. Of these two groups the low Cl value of the anthophyllite hornfelses is the most significant and indicates Cl migration during metasomatism.

Variation in F content of the Land's End aureole basic hornfelses with distance from the granite has been discussed previously by Floyd (1966). A general increase in the F/H_2O+ ratio as the contact is approached was interpreted as due to the increased F content of the primary hydrous phases. A similar increase in the associated pelitic rocks has also been recorded (Bowler 1959, Floyd 1967). Cl, on the other hand, shows a comparable spread of values for the meta-basic hornfelses within 80 metres of the contact (700 - 2000 p.p.m.) to those rocks situated about 800 metres away (700 - 1500 p.p.m.). Average values only show a small insignificant rise in Cl content as the contact is approached (from 1200 to 1300 p.p.m.). Similarly, the anthophyllite-bearing hornfelses with uniform low Cl show no change with distance. The cummingtonite-bearing hornfelses are the only group that show a steady progressive increase from 240 p.p.m. to about 1100 p.p.m. Cl. Again a wide scatter of values is seen near the contact (1000 - 1400 p.p.m.) for these rocks, although it is much smaller than that exhibited by the meta-basic parental rocks.

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TABLE 1. Average F and Cl in some metamorphosed Cornish greenstones compared with mean values of basic rocks.

Rock type	F ppm	Cl ppm	Number of samples for F & Cl resp.	Locality
Actinolite-bearing basic hornfelses	743	1207	8 each	Land's End granite aureole, Penwith peninsula, Cornwall.
Hornblende-bearing basic hornfelses	1087	1355	13 & 18	
Hornblende-bearing intermediate hornfelses	857	1253	8 & 9	
Anthophyllite-bearing magnesian hornfelses	1079	280	4 & 5	
Cummingtonite-bearing magnesian hornfelses	1039	945	15 each	
Calc-silicate-bearing calcareous hornfelses	1472	956	4 each	
Low-grade meta dolerite-diabase	198	162	14 each	Cudden Point, S.Cornwall coast.
Tholeiitic dolerite	392	76	34 each	Great Lake intrusion, Tasmania (Greenland and Lovering 1966).
Average basalt	360	—	130 —	**Compilation** (Fleischer and Robinson 1963)
Average gabbro/diabase	420	—	26 —	
Average basalt	400	60	— —	**Compilation** (Turekian and Wedepohl 1961)

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3. **Regional greenstones**

The Cudden Point greenstone is a tholeiitic basalt with a variably developed low-grade greenschist mineralogy (Floyd and Lees 1972). Compared with other basic rocks (Table 1) the greenstone has a lower F content and a smaller dispersion of values. The majority of the F probably resides in the hydrous phases (amphibole rather than chlorite ?) or in the few apatite crystals observed. Cl apparently behaves differently to F and shows higher values and a much wider range in the greenstone than in other non-metamorphosed basic rocks (Table 1). It is possible that this could be due to either sea-water contamination (Cudden Point being situated on the coast) or the presence of undetected liquid inclusions in some of the minerals. The averages listed in Table 1 exclude the two samples collected from the sheared contact zone as they contain much higher F and Cl (averaging about 700 p.p.m. F and 500 p.p.m. Cl) than the other samples. It is suggested that these elements may have been added to the greenstone margin, being initially derived from a subsurface ridge of the Godolphin granite. Other element additions (especially the alkalis) substantiate this observation (Floyd and Lees 1972).

4. F-Cl relationship

Figure 1 shows a plot of F vs. Cl for the regional and contact basic rocks. The markedly higher F and Cl content of the latter group is clearly demonstrated. Compared with the Great Lake tholeiite (Greenland and Lovering 1966) the Cl content of the regionally metamorphosed Cudden Point tholeiite is higher and also exhibits a much greater dispersion. As previously mentioned this could be due to sea-water contamination of the samples, although there appears to be a tendency for the high Cl samples to have correspondingly low F contents and vice versa. The Great Lake samples with higher F and lower Cl than the Cudden Point samples, accentuates this apparent trend. A similar trend is suggested by the decrease in Cl values from the low-grade contact basic hornfelses to the contact zone of the Cudden Point greenstone to the Wilmingstone greenstone MUM-1) just outside the Dartmoor granite aureole.

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The majority of the contact hornfelses show a different trend with a sympathetic and progressive increase of F with Cl. The anthophyllite-bearing metasomatic hornfelses are the exception and have probably lost some Cl during metasomatism.

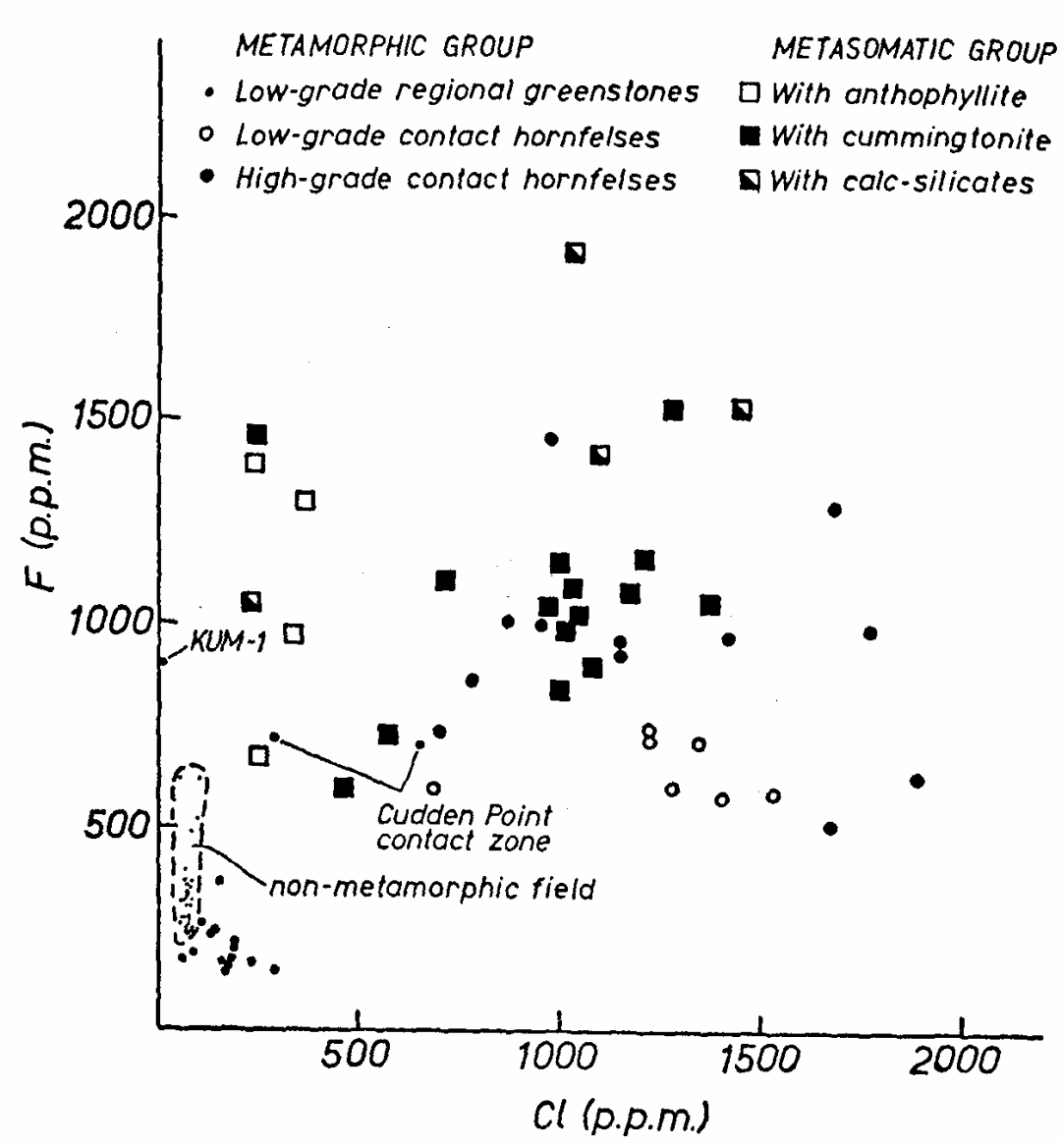

FIGURE 1. Plot of F against Cl for regionally metamorphosed greenstones and contact metamorphosed basic hornfelses from Cornwall. Various metasomatized hornfelses from the Land's End aureole have also been plotted. The enclosed non-metamorphosed tholeiite field is from Greenland and Lovering (1966).

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Ti-Zr CHARACTERIZATION OF SOME CORNISH PILLOW LAVAS

by P. A. Floyd and G. J. Lees

Abstract. Upper Devonian pillow lavas collected from Pentire Point, Kellan Head and Mullion Island, Cornwall, have been analysed for Ti and Zr. The absolute concentrations of these elements and their average Ti/Zr ratio indicate that the pillow lavas are alkali basalts with possible continental affinities. They also show similar Ti and Zr distributions to the associated intrusive greenstones of Cornubia.

The apparent increase in Ti and Zr from core to rim of some individually analysed pillows may be due to the dilution effect of extensive secondary calcite infilling the highly vesicular cores.

1. Introduction

Geochemical studies on modern volcanic rocks has added considerably to our knowledge of the chemical characteristics of basaltic rocks found in different tectonic environments. Basalts sampled from the ocean floor, island arcs and continental areas often have a distinct chemistry which can help in understanding the palaeoenvironment of ancient basaltic rocks. This is particularly important in the context of plate tectonics with the identification of ancient ocean floor basalts or volcanic rocks formed at active oceanic-continental margins.

Recent reconstructions of Armorican plate margins have suggested the presence of a subduction zone through or to the south of Cornubia during the Upper Palaeozoic (Johnson 1971 ; Hawkes 1971 ; Burrett 1972), while other models have placed this zone in the Tethyan region (Nicolas 1972 ; Floyd 1972a). In most cases little attention has been given to the chemistry of the Cornubian volcanics (intrusive greenstones and pillow lavas), although Floyd (1972b) demonstrated that the intrusive greenstones had alkali basalt characteristics (with possibly continental affinities). A geochemical study of the pillow lavas is therefore pertinent to see if they have similarities to the intrusives and also whether they are typical of the oceanic crust environment or not.

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One of the main problems of analysing submarine volcanics that have been deeply buried in a thick sedimentary pile, is that they are often affected by secondary processes which alter their initial chemistry. Late deuteric alteration, devitrification, submarine weathering and low-grade regional metamorphism severely change both major and trace element compositions (Hart 1969 ; Cann 1969 ; Hattori *et al.* 1972). However, studies on ocean floor basalts (Cann 1970 ; Pearce and Cann 1971). have indicated that the more refractory elements, such as Ti, Y and Zr, may be used to discriminate parental associations and characterize different basaltic types.

2. **Cornish pillow lavas sampled**

Some 23 samples of pillow lavas were collected and analysed for Ti and Zr by X.R.F., using International basic and ultrabasic rock Standards for calibration purposes. In view of the mineralogical and chemical variation often seen between core and rim of individual pillows, both the inner and outer parts of 6 pillow lavas were analysed separately.

Pillow lavas were collected from Pentire Point and Kellan Head (near Portquin) on the N. Cornish coast, and Mullion Island off the Lizard peninsula in the south. The Pentire Pillow Lava Group, which contains about 200 m. of lava, has been dated as L. Frasnian in age (Gauss and House 1972). Similarly the Mullion Island lavas are designated as Frasnian (Hendricks *et al.* 1971) on the presence of conodonts in associated calcareous horizons. The Kellan Head lavas may well be Upper Devonian also.

All the samples show varying degrees of secondary alteration to hydrous phases (chlorite and epidote) and have developed a typical low-grade regional metamorphic spilitic assemblage. Concentric zones of vesicules are common (especially at Pentire Point) and often infilled by either calcite or chlorite. Some pillows (not sampled) have either chert or crystalline calcite centres. Details of the mineralogy and geological setting of these rocks may be found in the I.G.S. Survey memoirs covering the areas sampled.

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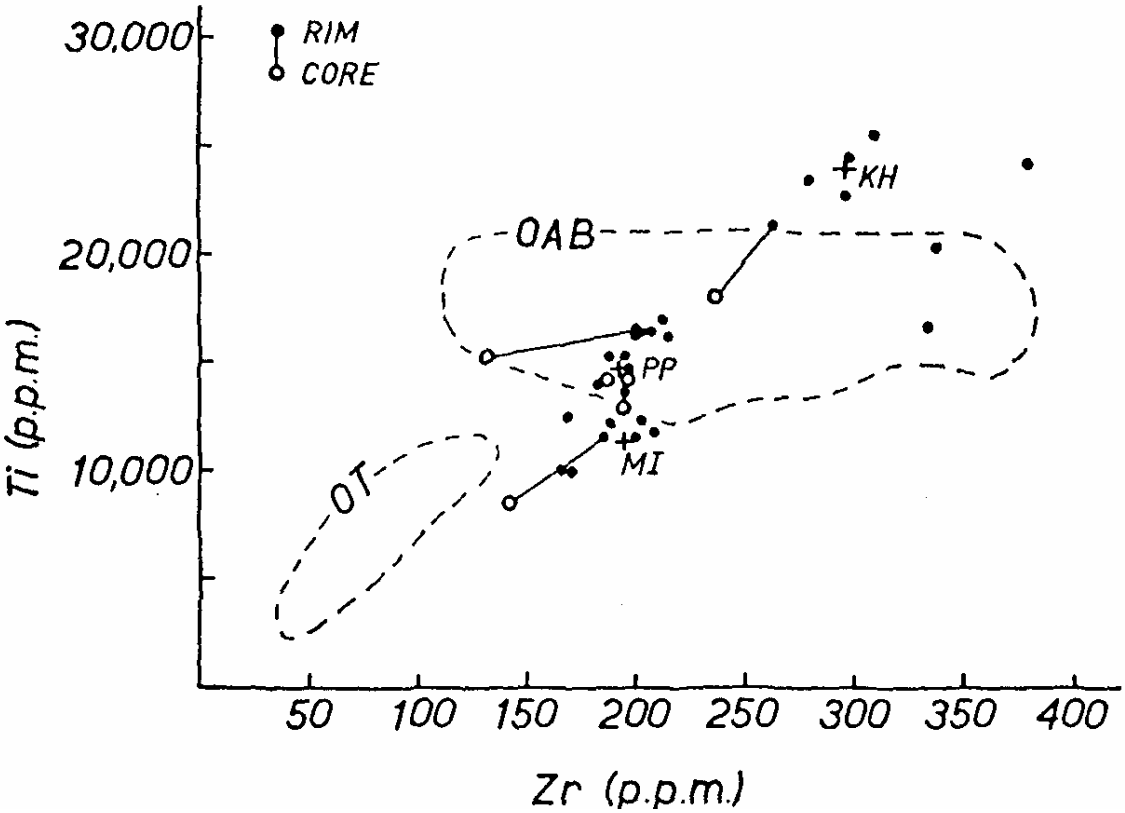

FIGURE 1. Ti-Zr plot for some Cornish pillow lavas. Lines join the analysed rim and core of separate pillows. Averages (crosses) for the three sampling areas are also shown : Pentire Point (PP), Kellan Head (KH) and Mullion Island (MI). The plot also includes 4 samples from Cosgrove (1972) and Bloxam and Lewis (1972).

The oceanic tholeiite (OT) and oceanic alkali basalt (OAB) fields are from Pearce and Cann (1971) and Floyd (1972b) respectively.

3. **Ti-Zr relationships**

Figure 1 shows a plot of Ti vs. Zr for the pillow lavas analysed plus 4 other samples taken from Cosgrove (1972, personal communication) and Bloxam and Lewis (1972). In all cases the Cornish pillow lavas have high Ti and Zr values and are clearly distinct from oceanic tholeiites in this respect. The Ti and Zr contents are typical of alkali basalts in general and show a close association with the intrusive greenstones which have been similarly designated (Floyd 1972b).

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The Cornish spilitic greenstones (both instrusives and pillow lavas) are thus predominantly alkali basalts and the question arises as to whether they are oceanic or continental in character. Certainly such a high proportion of alkali basalt material in an oceanic crust environment (where the majority of the rocks are tholeiitic) is unusual. In Figure 1 the majority of the pillow lavas fall in the oceanic alkali basalt field as plotted, although there is considerable overlap with continental analogues. In general continental alkali basalts have somewhat lower Ti contents, not unlike the lavas plotted here. Table 1 shows the average Ti / Zr ratios (in p.p.m.) for various continental and oceanic basaltic types compared with the Cornish greenstones. There is reasonable agreement between the Ti/Zr ratios for average continental alkali basalts (70) and both the intrusive greenstones (64) and pillow lavas (74).

TABLE 1. Comparison of average Ti/Zr ratios in Cornish greenstones with various oceanic and continental basaltic types.

Basaltic rock types	TiO_2 wt. %	Ti p.p.m.	Zr p.p.m.	Ti/Zr ratio
Average intrusive greenstones	2.31	13,860	218	64
Average pillow lava	2.81	16,860	229	74
Pentire Point pillow lavas (20)	2.46	14,760	194	76
pillow cores only (6)	2.32	13,920	182	77
pillow rims only (6)	2.59	15,540	205	76
Kellan Head pillow lavas (4)	4.02	24,120	297	81
Mullion Island pillow lavas (5)	1.94	11,640	195	60
Oceanic tholeiite	1.49	8,940	95	94
Continental tholeiite	1.20	7,200	224	32
Oceanic alkali basalt				
E. Pacific Rise	2.87	17,220	333	52
Gough Island	2.94	17,619	176	100
Continental alkali basalt	2.20	13,200	190	70

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Another interesting feature (Fig. 1) is the very high Ti and Zr content of the Kellan Head pillows compared with the other two localities. This could be due to the Kellan Head samples being more differentiated than the others, although without supporting chemical data it is not possible to evaluate this suggestion.

Variation within pillows from core to rim shows a marked increase in both Ti and Zr for 3 of the 6 pillows analysed. Analyses by Vallance (1965) of a Mullion Island pillow showed a similar increase in TiO, from core (1.39 wt.%) to rim (2.44 wt.%). It appears that both Ti and Zr can be affected by secondary processes acting on the pillows and cause migration of these elements towards the margins. However, the 3 samples that showed the maximum core-rim variation all have the greatest development of secondary calcite infilling large vesicles in the core of the pillow. In other words, the apparent decrease from rim to core might be a dilution effect due to the extensively developed calcite. To test this possibility all rim-core samples were leached with dilute acetic acid and reanalysed for Ti and Zr. The preliminary results show increases in absolute amounts of Ti and Zr for both rim and core compared with the initial untreated samples, although Ti and Zr concentrations now more closely correspond between rim and core. In some cases there is a marginal increase in both Ti and Zr from rim to core, although the results are not analytically significant.

ACKNOWLEDGEMENTS. Thanks are due to Dr. M. Cosgrove for the use of unpublished pillow lava analyses. Financial assistance for field work was gratefully received from the University of Keele Research Fund.

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GEOPHYSICAL SURVEYS OVER TWO ORE-BODIES NEAR TAVISTOCK, DEVON

by E. M. Durrance

Abstract. The results of Magnetic Total Field, Spontaneous Potential, Apparent Resistivity and Electro-Magnetic surveys over two sulphide bearing ore bodies near Tavistock show close agreement. A method for the quantitative analyses of spontaneous potentials to determine the limiting depth to the top of a polarised body is described.

1. Introduction

The geology of the Tavistock area of Devon consists of recumbently folded Upper Devonian and Carboniferous strata cut by the Dartmoor Granite and into which have been emplaced a series of metalliferous veins (Dearman and Butcher 1959). The Upper Devonian Whitchurch Green Slates, and occupy the southern margin of the area, while Lower Carboniferous siliceous slates and greenstones, and Upper Carboniferous turbidite sandstones and shales, form most of the area immediately to the north. The boundary between the Whitchurch Green Slates and the Carboniferous strata is considered by Selwood (1971) to be the westward continuation of the Holne thrust. After the emplacement of the Dartmoor Granite, which forms the eastern margin of the area, faulting and mineralisation occurred, predominantly along E-W and N-S lines. The location and trend of the major lodes together with their most productive metals is given by Dines (1956).

Preliminary geophysical surveys of this metalliferous region located a number of interesting anomalies, two of which were chosen for a more detailed investigation to compare the response of different geophysical techniques. The first of these (Area 1) was an ENE trending spontaneous potential anomaly of some 350mV to the north-east of Mary Tavy ; the second (Area 2) a N-S spontaneous potential anomaly of about 300mV, east of Tavistock.

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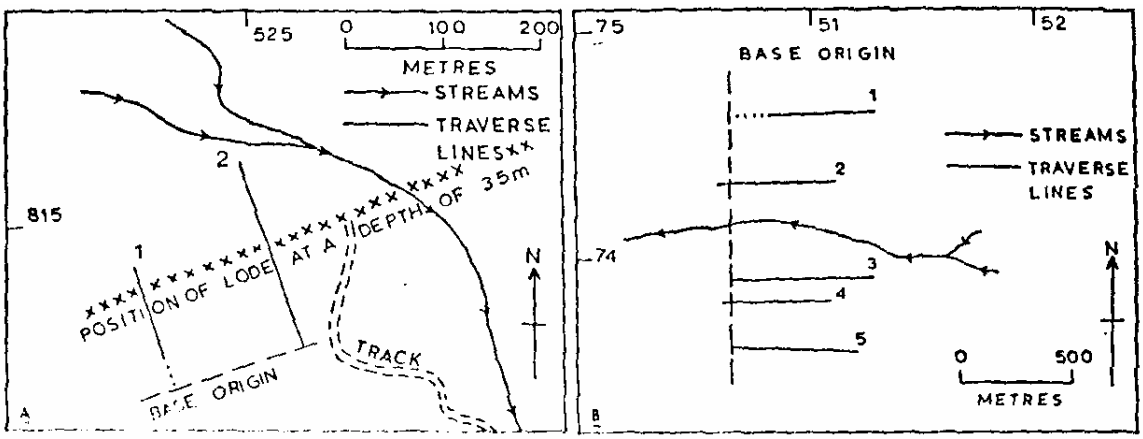

FIGURE 1.

A. Geophysical Traverses in Area 1. B. Geophysical Traverses in Area 2.

In Area 1 two traverse lines (Fig. 1A) were established, crossing the trend of a known lode approximately at right angles. Traverse I was surveyed using only a spontaneous potential direct field method, while traverse 2 received surveys for spontaneous potential, apparent resistivity, magnetic total field, and three electro-magnetic configurations, namely : moving transmitter-receiver horizontal coplanar and vertical coaxial coils, and a very low frequency (radio-wave) receiver system.

In Area 2 five traverse lines were surveyed (Fig. 1B) each trending approximately E-W. Lines 1, 2, 4 and 5 were spontaneous potential direct field surveys, but line 3 was surveyed using spontaneous potential, apparent resistivity, magnetic total field and two electro-magnetic configurations: moving transmitter-receiver horizontal coplanar coils, and the very low frequency (radio-wave) receiver system.

2. Equipment and Technique

(A) Spontaneous Potential:

Geonics S.P. 19 high impedance voltmeter. Copper electrodes in copper sulphate solution. Direct field readings, one fixed electrode.

Area 1: (traverses 1 and 2) spacing 10m.

Area 2: (traverses 1, 2, 4, 5) spacing 20m ;(traverse 3) spacing 20m, 10m near minimum.

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(B) Magnetic :
Elsec Proton Magnetometer. Total field.
Area 1: (traverse 2) spacing lOm.
Area 2: (traverse 3) spacing 20m, lOm near anomaly.

(C) Apparent Resistivity :
Nash and Thompson Tellohm. Wenner configuration.
Area l : (traverse 2) 10m spacing, 10m advance.
Area 2: (traverse 3) 10m spacing, 10m advance.

(D) Very Low Frequency (Radio-Wave) Electro-Mangetic:
Geonics E.M. 16.
Area 1: (traverse2) Transmitter NAA-Cutler, Maine at 17.8 kHz. Spacing 10m.
Area 2: (traverse 3) Transmitter GBR-Rugby, England at 16 6Hz. Spacing 20m.

(E) Moving Transmitter-Receiver Horizontal Coplanar Coils :
Geonics E.M. 17.
Area 1: (traverse 2) 1.6 kHz, seperation 30.5m spacing 10m.
Area 2: (traverse 3) 1.6 kHz, separation 30.5m, spacing 20m.

(F) Moving Transmitter-Receiver Vertical Coaxial Coils :
Geonics E.M. 17.
Area 1: (traverse 2) 1.6 kHz, separation 30.5m, spacing 10m.

3. Analytical Methods

Consider a vertical dipole as a source of spontaneous potential, with its lower pole at infinity, as represented by a point current source (I) at a depth (d) below the earth's surface. This will give rise to an image (I), assuming the resistivity of air to be infinity, at a height (d) above the Earth's surface. The potential difference (ΔVx) due to the source and its image between a point on the earth's surface vertically above the source, and a second point a horizontal distance (x) away, is then given by the function

$$\Delta Vx = \frac{2\rho I}{4\pi}\left(\frac{1}{d} - \frac{1}{\left(d^2 + x^2\right)^{\frac{1}{2}}}\right)$$

where ρ is the resistivity of the ground (Heiland 1946).

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But $Vo = \frac{2\rho I}{4\pi d}$

Where Vo is the potential at the point on the earth's surface vertically above the source,

therefore $\Delta Vx = Vo.d\left(\frac{1}{d} - \frac{1}{\left(d^2 + x^2\right)^{\frac{1}{2}}}\right)$

and $d = \left(\frac{x^2(Vo - \Delta Vx)^2}{Vo^2 - (Vo - \Delta Vx)^2}\right)^{\frac{1}{2}}$

but $Vx = Vo - \Delta Vx$

where Vx is the potential at the second point,

therefore $d = \left(\frac{x^2.Vx^2}{Vo^2 - Vx^2}\right)^{\frac{1}{2}}$

Sulphide ore-bodies acting as sources of spontaneous potential are, however, unlikely to possess either current source or sink at the extremities of their mineralisation. Also, even for near vertical bodies, the lower pole is unlikely to be at infinite depth. Both factors act to reduce the real value of the depth to the top of the ore-body from that yielded by use of this equation. A better presentation would therefore be to consider that the value of (d) obtained by this equation is the maximum depth at which the top of the ore-body is likely to be found,

that is $z \leq \left(\frac{x^2.Vx^2}{Vo^2 - Vx^2}\right)^{\frac{1}{2}}$

where z is the depth to the top of the ore-body.

The purpose of carrying out a number of different geophysical surveys over the ore-bodies presumed to produce the spontaneous potential anomalies mentioned above, was to test the use of this relationship in comparison with that of conventional analysis of other geophysical anomalies.

Spontaneous potential anomalies may also be analysed according to techniques described by Heiland (1946) and Meiser (1962) which yield approximations of the depth to the centre of the polarised body. Similarly, magnetic anomalies may be analysed in a number of ways. The limiting depth to the top of a magnetic body is given by methods described by Smith (1959) :

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Parasnis (1963) gives a method for approximating the depth to the top of a magnetic body, and Bruckshaw and Kunaratnam (1963) give an absolute method for determining this value and the width for dyke-like magnetic bodies. A method by Werner, described by Parasnis (1966), gives an approximation of the depth to the top of the body and its dip. The quantitative analysis of very low frequency (radio-wave) electro-magnetic anamalies is still at an early stage of development, but techniques have been developed which yield the depth to the centre of the conducting body producing the anomaly (Geonics 1969) and which appear to have been successfully applied. The analysis of anomalies produced by coplanar and coaxial coil, moving transmitter-receiver systems is described by Parasnis (1966) and Grant and West (1965), to obtain information about the position, depth and conductivity of any body producing an anomaly. In contrast, however, the analysis of apparent resistivity anomalies produced by constant separation traverses apart from indicating the position of zones of high and low conductivity does not easily yield more quantitative information.

4. Results

(A) Area 1

Spontaneous potentials recorded on traverse 1 (graph G, fig. 2) show a distinct minimum of about -350mV 100m north-west of the base origin. Spontaneous potentials recorded on traverse 2 (graph A, Fig. 2) indicate, however, two overlapping anomalies, one (designated 'a') of -250mV, 50m north-west of base origin, the other (designated 'b') of -180mV, 100m north-west of base origin. The quantitative analyses of these anomalies, based upon the recognition of the anomaly in traverse 1 as corresponding to that of anomaly 'b' in traverse 2, are given in Table 1.

A negative magnetic anomaly of approximately 120 y occurs about 50m north-west of base origin on traverse 2 (graph B, fig. 2). The quantitative analysis of this anomaly is similarly given in Table 1. The very low frequency (radio-wave) receiver system produced a distinct anomaly 100m north-west of base origin on traverse 2 in both Real (-28% to +28%) and Imaginary (- 10% to +5%) components (graph E, fig. 2). The analysis of this anomaly is also given in Table l.

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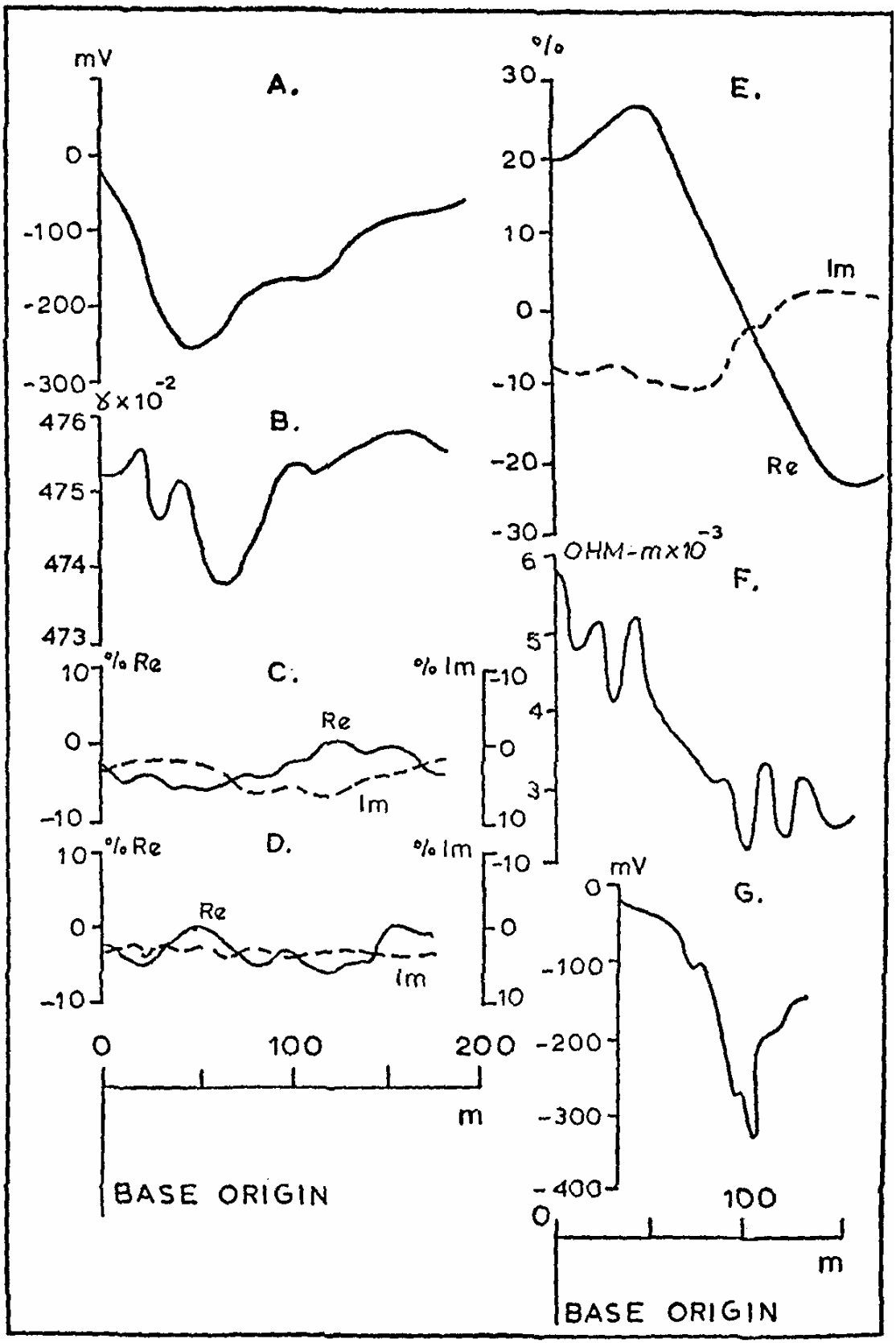

FIGURE 2. Results of Geophysical Traverses in Area 1. Headings A-F relate to traverse 2, heading G to traverse 1.
A. Spontaneous Potential ; B. Magnetic total field.
C. Electro-magnetic, vertical coaxial coils.
D. Electro-magnetic, horizontal coplanar coils
E. Very Low Frequency (radio-wave) electro-magnetic.
F. Apparent Resistivity ; G. Spontaneous Potential.

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TABLE 1

Analyses of Geophysical Anomalies in Area 1
Depths in metres to centre and top of sheet

Traverse No.	Character	Anomaly	Centre	Top	Method
1	Spontaneous Potential	-	34	≤ 7	-
2	,, ,,	a	35	≤11	-
2	,, ,,	b	75	≤ 9	-
2	Magnetic	a	-	≤18	Smith, 1959
2	,,	a	-	15	Parasnis, 1963
2	V.L.F. (radio-wave) E.M.	b	100	-	-

Apparent resistivity along traverse 2 (graph F, Fig. 2) indicates a fluctuating decrease over a distance of 100m from base origin, but no further analysis of this data appears justifiable. Similarly, the results of the moving transmitter-receiver electro-magnetic surveys in both horizontal coplanar and vertical coaxial coil modes (fig. 2, graphs D and C) do not show characteristics capable of quantitative analysis.

(B) Area 2

A spontaneous potential anomaly of -350mV occurs on traverse 1 at 300m from base origin, of -280mV at 290m on traverse 2, of -390mV at 260m on traverse 3, of -320mV at 260m on traverse 4 and an indistinct -100mV between 200m and 300m on traverse 5. As base origin is a line striking due north, the presence of an anomaly trending a few points east of north and closing to the south is clearly indicated by these results (shown in detail in Fig. 5). The quantitative analysis of the anomaly shown by traverse 3 is given in Table 2.

Magnetic field strength recorded along traverse 3 is given as graph A in Fig. 6. This shows a positive anomaly of about 26 γ approximately 250m from base origin. On removal of the effects of a west to east regional increase of 30 γ per kilometre, and after smoothing, this anomaly shows a component of -6 γ at 210m from base origin and a component of 24 γ, 250m from base origin. The results of quantitative analysis of this anomaly are given in Table 2.

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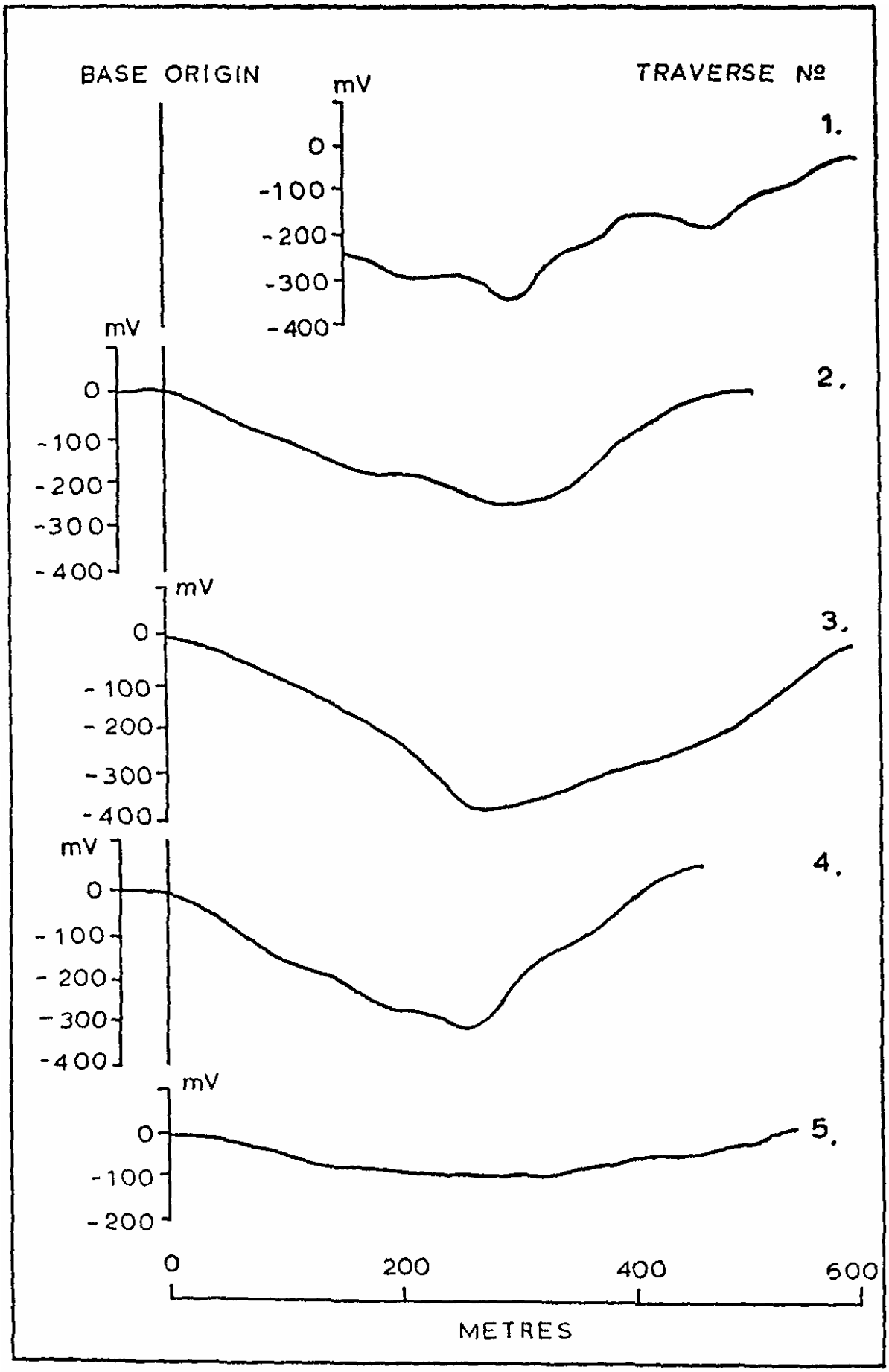

FIGURE 3. Results of Spontaneous Potential traverses in Area 2.

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TABLE 2.

Analyses of Geophysical Anomalies recorded on Traverse 3 in Area 2
Depths in metres to centre and top of sheet

Character	Centre	Top	Method
Spontaneous Potential	200	≤28	-
Magnetic	-	≤26	Smith, 1959
,,	-	20	Parasnis, 1963
,,	-	20	Parasnis, 1966
,,	-	18	Bruckshaw & Kunaratnam, 1963
V.L.F. (radio-wave) E.M.	220	-	-

Apparent resistivity along traverse 3 shows a fluctuating level of approximately 4000 ohm-m at each end, but a distinct anomaly occurs at about 250m from base origin where the value has fallen to only 1200 ohm-m. The form of the anomaly is asymmetrical, with the steepest gradient on the western margin. The profile is shown as graph D in Fig. 4 but apart from the coincidence in position found between this and other anomalies recorded on traverse 3 no further analysis is possible.

Of the two electro-magnetic configurations used for surveying traverse 3, the moving transmitter-receiver horizontal coplanar coil system produced no distinct anomalies, and the results (graph B, fig. 4) are not capable of further analysis. The very low frequency (radio-wave) receiver system produced, however, a distinct anomaly about 260m from base origin in both Real (-12% to +10%) and Imaginary (-2% to + 12%) components. The quantitative analysis of these results (graph C, Fig. 4) is given in Table 2.

5. Geological Interpretation

(A) Area 1

Spontaneous potentials of the magnitude recorded in this area are characteristic of sulphide mineralisation. The values found on traverse 2 indicate the presence of two sources of anomaly, the more southerly possessing the greatest influence. On traverse 1 the southerly anomaly is negligible, while the more northerly has increased in magnitude when compared with the equivalent position on traverse 2. This suggests that the lode is not a single sheet but consists of at least two ore-bearing veins

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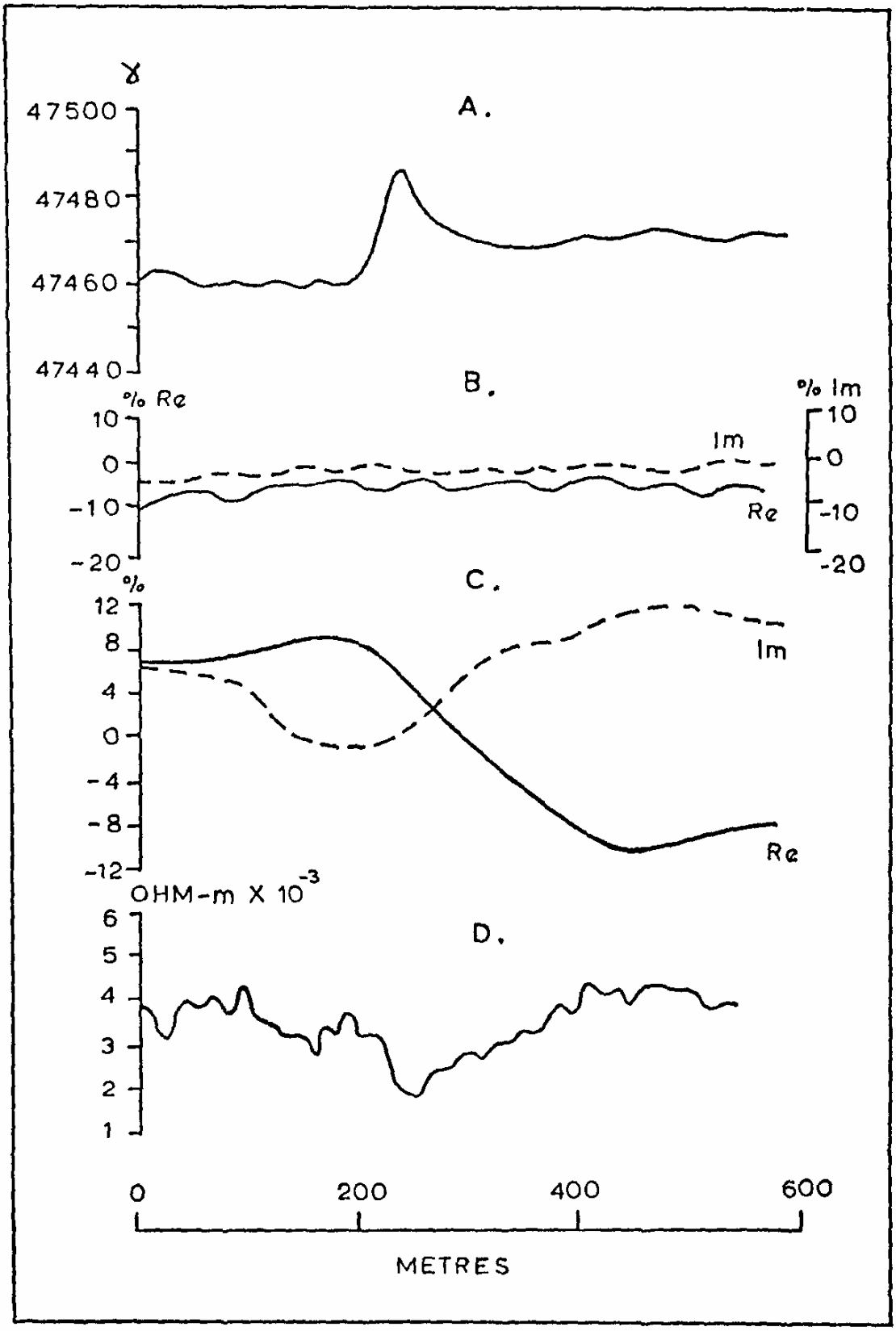

FIGURE 4. Results of Geophysical surveys along Traverse 3, Area 2.
A. Magnetic total field.
B. Electro-magnetic, horizontal coplanar coils.
C. Very Low Frequency (radio-wave) electro-magnet.
D. Apparent Resistivity.

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The coincidence in position between different geophysical indications on traverse 2 is pleasing to note. A magnetic total field anomaly lies within 5m of the spontaneous potential minimum, an agreement to be expected as both indicate upper pole positions. The very low frequency (radio-wave) electro-magnetic and apparent resistivity anomalies are however, displaced to the north-west of this position. As these are sheet effect anomalies a dip of the lode toward the NNW is therefore implied. Quantitative analyses of the electro-magnetic anomaly yields a depth to the centre of the sheet of about 100m, as the anomaly is displaced by approximately 50m horizontally, a dip of about 63º is indicated. This value agrees favourably with the known attitude of the lode, but it is interesting to note that no indication of this dip is given by the magnetic anomaly alone.

The geophysical anomalies also agree with the known mineralogy of the lode. The presence of arsenopyrite would account for the large spontaneous potential anomaly while the additional presence of pyrite would produce a sufficient conductivity contrast to account for the very low frequency (radiowave) electro-magnetic and apparent resistivity anomalies. Some pyrrhotite is also probably present (with a reversed remnant magnetisation) to account for the magnetic anomaly. The lode is recorded as possessing an average width of about 2m, which also agrees with the character of the magnetic anomaly.

(B) Area 2

The magnitudes of the spontaneous potential anomalies recorded in this area again indicate the presence of sulphide mineralisation. The N-S trend of the anomaly is very clear, although the reduction in magnitude found on traverse 5 shows that the lode producing the anomaly dies out toward the south. Only a single source of anomaly is indicated by the results of all spontaneous potential traverses. As in Area 1, there is broad agreement between the position of the various geophysical indications found on traverse 3. Apparent resistivity, spontaneous potential, magnetic total field and very low frequency (radio-wave) electro-magnetic anomalies all occur about 260m from base origin. This coincidence suggests that the body producing the anomalies has a near vertical dip. Quantitative analysis of the

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magnetic anomaly yields in fact a dip of 74° to the east. The same analysis indicates, however, that the body has a width of only approximately 2m, while the associated apparent resistivity anomaly occupies a zone about 100m wide. This discrepancy probably indicates that the apparent resistivity anomaly is produced by a fault zone, and that the mineralisation producing the other geophysical anomalies occupies only a narrow part of the zone. The asymmetry of the apparent resistivity anomaly could however, still be explained in terms of the steep easterly dip of the body producing the magnetic anomaly lying parallel to that of the fault zone producing the apparent resistivity anomaly. Although no fault is shown in the position of this anomaly by Dearman and Butcher (1959), they do indicate the presence of faulting in the area east of Tavistock with an approximately N-S trend.

As the majority of the N-S lodes in the Tavistock area are lead bearing, it is probable that the anomalies found on traverses 1 to 5 lie over a body containing galena. The general character of the geophysical anomalies support this conclusion and suggest, in addition, the likely presence of arsenopyrite and pyrrhotite.

6. Conclusions

All geophysical surveys, with the exception of the moving transmitter-receiver horizontal and vertical coil systems, obtained distinctive anomalies in both Area 1 and Area 2. The failure of the moving transmitter-receiver electro-magnetic surveys suggests that either the conductivity contrast between the ore-bodies and the country-rock is too small to be detected by these methods, or, more likely, that too short a coil separation was used.

Analysis of the spontaneous potential anomalies by the methods of Heiland (1946) and Meiser (1962) give positions for the centre of the dipole which are consistently more shallow than those for the position of the sheet centre obtained from analysis of the very low frequency (radio-wave) electro-magnetic anomalies, namely : 75m: 100m for Area 1, traverse 2 anomaly ‘b’; and 200m: 220m for Area 2, traverse 3. Variations as great as 250% are therefore found. For these steeply dipping sheets the method to obtain a limiting depth to the top of a

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body producing a spontaneous potential anomaly yields, however, results which are more consistent with those obtained from other geophysical indications, namely: ≤llm for Area 1, traverse 2 anomaly 'a,' compared with ≤18m (after Smith 1959) and approximately 15m (after Parasnis, 1963) for analysis of the associated magnetic anomaly. Similarly, for Area 2, traverse 3, analysis of the spontaneous potential anomaly indicates a depth of ≤28m, compared with ≤26m (after Smith 1959), about 20m (after Parasnis 1963), 18m (Bruckshaw and Kunaratnam, 1963) and approximately 20m (after Parasnis 1966) derived by analysis of the associated magnetic anomaly.

ACKNOWLEDGEMENTS. Sincere thanks are offered to Dr. M. Stone and Mr. S. R. Haygreen for their help in conducting the geophysical surveys, and the Mr. S. R. Haygreen and Mr. G. R. Colley for carrying out some of the preliminary spontaneous potential surveys. Grateful thanks are also offered to Mr. P. J. Fenning and Mr. R. Henderson of Barringer Research for the loan of the Geonics E.M. 16 and E.M. 17 equipment.

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The sequence of emplacement of basic dykes in the Lizard Complex, South Cornwall (Abstract): by Alan Victor Bromley.

Relatively undeformed basic dykes are known to occur in peridotite, gabbro and Traboe Hornblende Schists of the Lizard Complex. In the area immediately south of Porthoustock basic dykes emplaced into the gabbro are exceedingly abundant and often make up more than 50% of the continuous coastal exposures. In this area, three groups of fine grained basic rocks occur in the gabbro.

1. Sheets and lensoid masses of fine grained olivine-clinopyroxene-plagioclase rock with pronounced xenoblastic texture. These may be early deformed basic dykes but are more probably beerbachites, xenoliths of fine grained basic rocks engulfed by the gabbro and metamorphosed under pyroxene-hornfels facies conditions.

2. Plagioclase porphyritic and xenocrystic meta-basalts which trend approximately NW-SE and dip at anglyes up to 45' in either direction. These dykes are commonly intruded by leucocratic hornblende diorites and microgranites which form net veined complexes within the dykes and in the adjacent gabbro.

3. NW-SE trending vertical dykes of non- or sparselyporphyritic amphibolite which are later than types (1) and (2) above. Sometimes two or more dykes of this type occupy the same fissure without evidence of chilling at their mutual contacts.

The pattern of dyke distribution near Porthoustock, coupled with evidence drawn from elsewhere in the Lizard Peninsula, suggests that the rocks of that area may be a remnant of the root zone of a sheeted diabase complex.

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STRUCTURAL ZONES IN VARISCAN PEMBROKESHIRE

by Paul L. Hancock

Abstract. Four structural zones can be recognised within the deformed Upper Palaeozoic sedimentary rocks at the northern margin of the Variscan fold belt in Pembrokeshire. Differences in style between the zones are attributed to the direction of underlying structural trends, northwardly declining stress differences, lithological contrasts, and the partially independent tectonic evolution of fault blocks.

1. Introduction

Within the deformed Upper Palaeozoic sedimentary rocks of south Pembrokeshire four structural zones can be recognised. This paper outlines the structural style of each zone and offers some reasons for the differences between them. It is hoped that the account will extend the idea of structural zones from South West England (see Dearman *et al.* 1971) to the northern margin of the Variscan fold belt in South West Wales. The principal sources of information apart from those cited in the text have been the Institute of Geological Sciences 1: 10560 and 1:63360 maps and the author's own observations.

The two major zones in Pembrokeshire are distinguished on the basis of axial trend, and each minor zone is characterised by its own assemblage of structural characteristics such as fold shape, size and vergence, the geometry of the fault system, and the relative development of cleavage and shear zones containing en-echelon veins. The terms macroscopic and mesoscopic fold are employed to describe folds of greater than 250 m and less than 250 m axial plane separation respectively. Compared with South West England there are few folds on the scale of a hand specimen in south Pembrokeshire.

2. Outline structural history

It would be inappropriate to review', here in detail the stratigraphic and tectonic evolution of Pembrokeshire during the Palaeozoic. The oldest movements are those of the Caledonian

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orogeny. North of the Benton Fault (Fig. 1), the Precambrian and Lower Palaeozoic rocks are overlain with angular unconformity by the Upper Palaeozoic cover. The alignment of the principal Caledonian folds is generally WSW except near Haverfordwest, where it is W. Upper Palaeozoic rocks south of the Benton Fault generally succeed the Lower Palaeozoic without a marked angular unconformity. Some restricted belts, which later developed into major strike faults, were nevertheless probably active south of the main Lower Palaeozoic outcrop throughout much of Lower Palaeozoic time.

Overstep and overlap plus thickness and facies changes in the Upper Palaeozoic sequence provide evidence for continuing instability across the major fault lines (Sullivan 1965, 1966 ; Owen 1971 ; Sanzen-Baker 1972). A few widespread horizons of sedimentary slump folds in the Carboniferous could indicate intermittent seismic shuddering on the proto-faults (Kelling & Williams 1966 ; Jenkins 1962). The main Variscan movements which produced most of the visible tectonic deformation were late Carboniferous or early Permian events.

Hancock (1964) has argued, using evidence from gash-breccias of presumed Triassic age (Dixon 1921), that the joints in the Upper Palaeozoic rocks of Pembrokeshire developed before the breccias. If the presumption on age is correct it follows that the folds and faults of the region, which pre-date the joints, are entirely a product of the Variscan movements. However, Thomas (1971) has recently claimed that some of the breccias are of mid-Tertiary age and related to movements on faults. Thus there may have been shearing of Alpine age on some fractures as in South West England (Dearman 1963).

3. **Description of the structural zones**

The primary division of the county is into Caledonian Pembrokeshire in the north and Variscan Pembrokeshire in the south. The subdivision of Variscan Pembrokeshire into Northern and Southern Systems, with WSW and WNW trending folds respectively, was recognised by Strahan (in Strahan *et al.* 1914) and followed by Dunning (1966). Here the Southern System is called Zone I and the northern System is called Zone II. Zone I is further divided into minor zones (Fig. 1).

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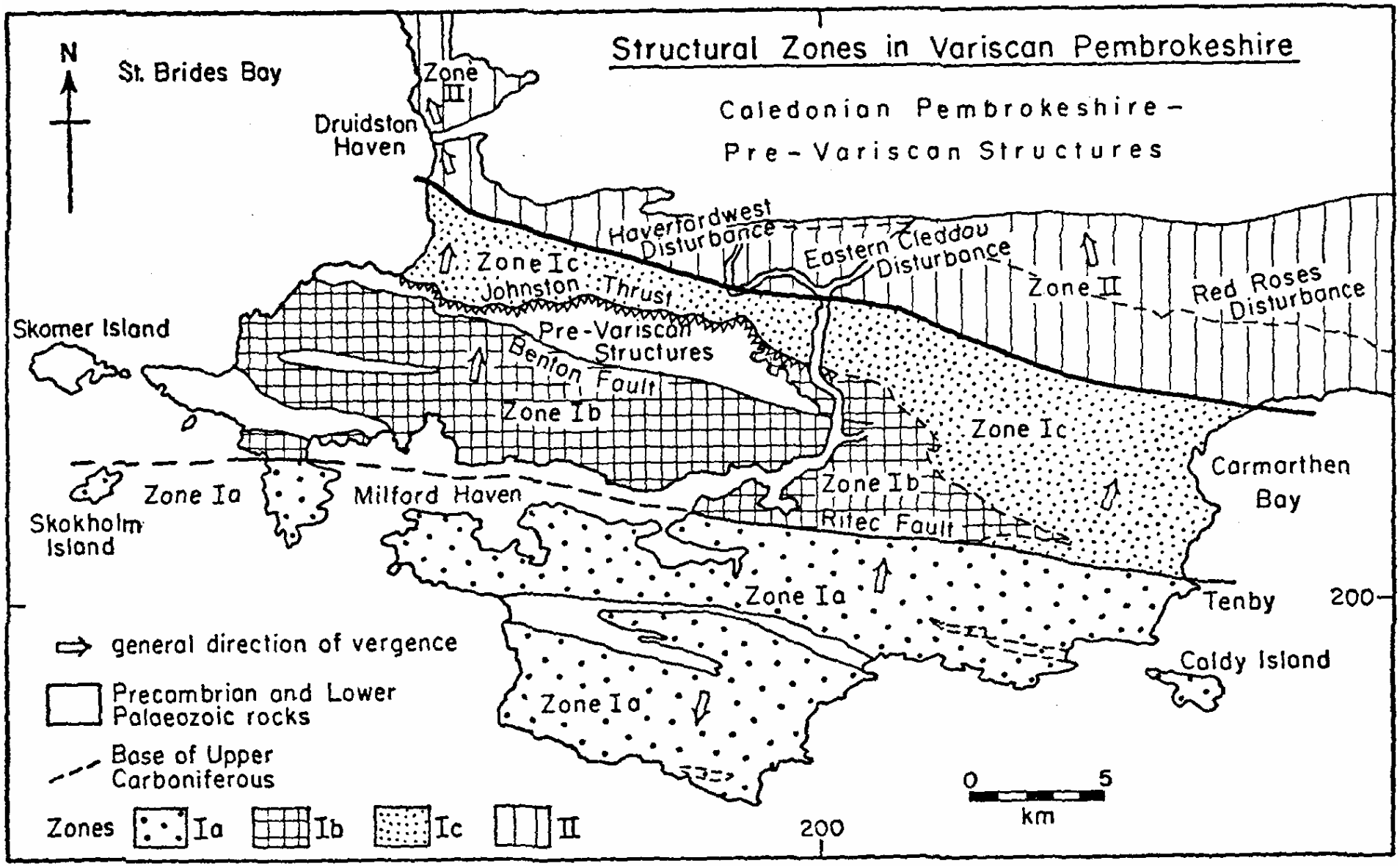

FIGURE 1. Structural Zones in Variscan Pembrokeshire

(a) Zone la

The outcrop of the high-angle Ritec Fault marks the northern boundary of the zone. The zone is characterised by large asymmetric macro-folds, some of them periclinal, trending WNW. At the surface they principally affect the Old Red Sandstone and the Lower Carboniferous limestones, but Lower Palaeozoic rocks occur in the cores of some anticlines, and Upper Carboniferous rocks in two synclinal outliers.

Relative to the centre line of the principal structure, the compound Orielton anticline, there is in the eastern part of the Pembroke peninsula a convergent-downward fan of steeply inclined axial planes. Zone la is the only zone containing a well-defined belt of south verging folds in addition to generally north verging folds.

The plunge of most macro-folds is 5° to 15°, mainly to the east. There are only four important folds which entirely or partly plunge to the west. Folds generally replace each other obliquely along the strike but they do not appear to be part of a belt of én-echelon folds.

The axial plane separation of the larger macro-folds generally exceeds 1 km and some of them have axial extents of greater than 10 km. Smaller macro-folds can rarely be traced for more than 3 km along the strike and their axial planes are separated by 0.25 to 1.0 km. In contrast to Zone Ic and much of Zone II, parasitic meso-folds are relatively uncommon. They generally occur as either (1) congruous, inequant, isolated pairs of anticlinal and synclinal buckles on otherwise uniformly dipping macro-fold limbs, or (2) incongruous, nearly equant axial buckles in the cores of some macro-folds. The axial plane separation of most meso folds is less than 100 m.

Most macro- and meso-folds are inequant, the shorter limb generally being more steeply inclined that the longer limb. The inclinations of steep limbs are mainly from 65°, through 90° to inversion by up to l0°. Moderately dipping limbs are generally inclined between 35° and 45° and there are some relatively large areas of dip of less than 20°. Interlimb angles are generally between 30° and 130°. Compared with Zone Ic well-developed chevron folds are uncommon.

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South Pembrokeshire has become a classic example of the relationship between folds, thrusts and wrench faults (Anderson 1951). The complete fault pattern is developed only in Zone Ia. Two sets of strike faults, which are thrusts, cut the folds. North of the centre line of the Orielton anticline they overthrust mainly to the north, while to the south of the centre line overthrusting to the south is common. These are directions of transport which accord with the vergence of the folds.

The folds and thrusts are cut by two conjugate sets of nearly vertical wrench faults symmetrically arranged about the axial trend of the folds. A widespread, symmetrical, conjugate wrench fault system is largely missing in the zones north of the Ritec Fault. Faults of the NNW trending set are dextral and they are more numerous than members of the NE trending sinistral set. Both sinistral and dextral wrench faults are more abundant than thrusts. During the author's survey of minor wrench faults it was found that some conjugate pairs of adjacent minor faults are orientated so that the obtuse bisectrix between them is parallel to the axis of the fold in which they are contained, and the acute bisectrix is approximately perpendicular to the axial plane.

Cleavage style varies from a slaty cleavage in some mudstones in the west, to an imperfect spaced cleavage in many mudstones in the east and in some sandstones throughout the zone. The spaced cleavage is partly a fracture cleavage initiated by hydraulic fracturing according to Price and Hancock (1972). The change along the strike from a spaced to a slaty cleavage is not uniform, and some rocks are uncleaved. Cleavage surfaces intersect bedding planes along a lineation which is parallel to nearby fold axes, and they generally form a convergent fan within folds.

Conjugate shear zones containing arrays of en-echelon quartz or carbonate veins are abundant in some sandstones and limestones. Zones and veins are mainly perpendicular to the bedding and the obtuse bisectrix between conjugate zones is parallel to the b lineation formed by the cleavage on the bedding. Conjugate shear zones of this type are rare in the other zones except in the south western part of Zone Ib.

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(b) Zone Ib

The southern boundary of the zone is the outcrop of the Ritec Fault and the northern boundary is, in the west, the Benton Fault, and, in the east, the junction between the Lower and Upper Carboniferous. The principal fold, the Winsle-Carew anticline, exposes mainly the Old Red Sandstone but its core contains Silurian rocks, and there is a partial envelope of Carboniferous Limestone in the east. Approximately 30 km in length, the structure comprises the four separate eastward plunging anticlines of Winsle, Burton, Carew and Sageston.

The Old Red Sandstone rocks of the Winsle anticline are affected by many parasitic macro- and meso-folds mainly verging north and trending WNW but including a few WSW, NW and W trends. The Winsle anticline folds are not as markedly inequant or asymmetric as those in Zone Ia, and the inclinations of the steep limbs are generally less than in that zone. The Burton, Carew and Sageston anticlines are simpler, generally more open structures plunging east at angles of up to 30°. Their limbs are disturbed by some meso-folds, many of which on the Burton and Carew anticlines plunge steeply east.

South of the Winsle anticline the Musselwick Fault forms the northern margin of a south dipping block of Ordovician, Silurian and Old Red Sandstone sedimentary rocks and Silurian volcanics. Along the southern side of this block and on the southern limb of the Winsle-Carew anticline the general WNW strike of the beds is locally deflected by up to nearly 90° where there are folds plunging steeply to the south-east.

The only major strike faults within the zone are the Musselwick Fault and nearby parallel faults, and the fault along part of the northern limb of the Sageston anticline. Minor thrusts are abundant in the envelope of the Winsle anticline. There is no well-developed system of conjugate wrench faults symmetrically arranged about the axial trend of the folds as in Zone Ia. Some of the cross-faults undoubtedly have a component of strike-slip on them, but those which displace the Skomer Volcanic Group are interpreted as normal faults (Thomas, in Cantrill *et al*. 1916).

As in Zone Ia, there is a slaty cleavage in some mudstones of the Old Red Sandstone in the western part of the zone, and a

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spaced cleavage in mudstones in the east and in some sandstones throughout the zone. There are some tracts of nearly axial planar cleavage in the Winsle anticline.

Conjugate shear zones perpendicular to beds and containing en-echelon veins are rare except in some sandstones near to the top of the Silurian and base of the Old Red Sandstone in the extreme south west of the zone.

(c) Zone Ic

West of the Daugleddau the southern boundary of the zone is the outcrop of the Johnston Thrust, and east of the Daugleddau it is the northern boundary of Zones lb and Ia. The northern margin of the zone and the Southern System is an ill-defined line in the Coalfield drawn approximately along the northern limit of mainly WNW trending folds. The rocks of the zone are all of Upper Carboniferous age.

Most of the asymmetric folds plunge gently, trend WNW, and verge to the north. There are also a few WSW and W trending folds, and some which verge to the south. The abundance of inequant meso-folds, commonly of chevron shape, parasitic to mapped macro-folds is the distinguishing feature of most of the zone. In the southern part of the zone the folds are generally tighter and smaller, and some of their steep limbs are slightly overturned. A few otherwise nearly straight limbs are disturbed by gentle minor buckles with axial surfaces at high angles to the axial surface of the fold containing them. In the northern part of the zone the folds are generally more open except locally as on the limbs of a macro-anticline just south of Amroth, where there are tighter meso-folds, the axial surfaces of which are arranged in a strongly convergent fan on the macro-anticline.

Strike-faults are mainly reversed and generally overthrust to the north except where they are associated with south-verging folds. Most of them are high-angle faults but a few are low-angle. Some strike faults are curved, and some wholly or partially replace the steep limbs of folds. They probably indicate that folding and thrusting were partially synchronous. In the northern part of the zone some strike-faults show normal displacements.

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There is no well-developed symmetrical pattern of conjugate wrench faults as in Zone Ia. Some of the cross-faults show strike-slip displacement, but they are mainly at high angles to the axial trend of the folds, and many of them are probably tear faults partly synchronous with the folding. Slickensides on some cross-faults indicate that there was some oblique-slip.

There is a spaced cleavage in some sandstones throughout the zone but cleaved mudstones occur only in tile Southern part where their development is uneven. Well-developed conjugate shear zones containing en-echelon veins and perpendicular to the bedding are infrequent compared with Zones Ia and south-western Ib.

(d) Zone II

The southern boundary of the zone is the indefinite northern limit of Zone Ic, and its northern boundary is the outcrop of the unconformity between the Lower and Upper Palaeozoic rocks. The zone contains four relatively narrow WSW trending belts of folds and faults separated by less deformed rocks generally dipping to the south. The principal belts or deformation are, from east to west: the Red Roses Disturbance, the Eastern Cleddau Disturbance, the Haverfordwest Disturbance and the fault and fold zone which involves Lower Palaeozoic rocks at Druidston Haven. The disturbed belts are more widely spaced in the east than the west, so that on St. Brides Bay there is an almost continuous sequence of folds and faults, which near to the southern boundary of Zone II mainly differ in trend from those in Zone Ic. The principal effect of both the folds and the faults is to downthrow rocks to the north. Within and to the north of the main South Wales Coalfield, the Careg Cennan, Swansea Valley, and Vale of Neath Disturbances are other similarly orientated fold and fault belts (Owen 1971). The Pembrokeshire disturbances are sub-parallel to structural trends in nearby Lower Palaeozoic rocks, and they affect all the major divisions of the Upper Palaeozoic sequence.

The disturbances are comprised of folds and strike faults in belts up to 2 km in width and 20 km in lateral extent. Individual folds are generally impersistent, gently plunging and mainly trend WSW with subsidiary WNW and W alignments. North dipping limbs are usually shorter than south dipping limbs and

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many folds verge to the north, but some are nearly upright or southerly verging. The faults are mostly high-angle and downthrow to the north.

The structure of the Upper Carboniferous outlier of the Nolton-Newgale Coalfield is not known in detail. It is separated from the Little Haven-Amroth Coalfield by an upfaulted block of Ordovician rocks at Druidston Haven. The structures on St. Brides Bay are WSW trending relatively open folds. Inland there are mainly gentle or moderate dips to the west.

In Zone II there is no widely-developed system of conjugate wrench faults symmetrically arranged about the axial trend of the folds. Some cross-faults have a component of strike-slip along them but on others and many of the strike faults, there has been significant normal dip-slip. There is a weak cleavage in some siltstones of the Old Red Sandstone, and a spaced cleavage in some Upper Carboniferous sandstones. Conjugate shear zones containing en-echelon veins are developed parallel to some small normal or extension faults, but they do not generally form sets perpendicular to the bedding as in Zones la or south western Ib.

4. Discussion

The only worker since Strahan *et al.* (1914) to attempt a new tectonic subdivision has been Sanzen-Baker (1972), who, using stratigraphic and sedimentological evidence from the Silurian and early Old Red Sandstone, divides south Pembrokeshire into four allochthonous blocks and one autochthonous area. From south to north the blocks are separated by the Ritec, Musselwick and Benton Faults and the Johnston Thrust. Sanzen-Baker (1972, fig. 1) continues the Musselwick Fault east to join the Ritec Fault, and the Benton Fault and Johnston Thrust east through the Coalfield to Carmarthen Bay. The justification for extending eastwards all of these structures as major strike faults is not everywhere clear.

If it is accepted that successions which were once more widely separated have been brought together by faulting, many of the faults which are high-angle at the surface must, in order to accommodate the horizontal translation, either flatten out at

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depth or have a major component of transcurrent movement on them. Although Sanzen-Baker proposes the latter for the Musselwick and Ritec Faults, the former explanation probably applies to most of the major strike faults.

The differences in structural style between the zones probably reflect the influence of several controls, not all of equal importance, which at the time of deformation would have acted together, but which for convenience are listed separately here.

1. As Jones (in Strahan *et al.* 1914) realised, renewed movements on Caledonian directions underlying Zone II probably influenced the WSW alignment of the disturbed belts. In the southern parts of Pembrokeshire, where the Upper Palaeozoic sequence is thicker and there is not a widespread marked angular unconfonnity at its base, a WSW Caledonian direction may be absent or less pronounced, and it would be less effective.

2. The change from relatively tight to relatively open folds in Zone Ic suggests that from south to north there was possibly a decline in the magnitude of the differential stresses responsible for deformation.

3. In the Upper Carboniferous sequence of Zone Ic, thin alternating beds of sandstone and mudstone, of marked ductility contrast, and beneath relatively little cover, allowed the development of abundant meso-folds. The sequence of relatively competent, more deeply buried, Lower Carboniferous limestones of up to 2000m in thickness in Zones Ia and Ib probably inhibited the formation of many meso-folds. In the western parts of Zone II where the disturbed belts are more closely spaced the Upper Carboniferous rocks rest either on a thin Carboniferous Limestone sequence, or directly on the Lower Palaeozoic rocks of the Caledonian basement.

The structural style in Zone Ic is not unlike that in the Upper Carboniferous rocks in parts of the Mid-Devon Synclinorium. In both areas there are abundant small chevron folds, and the limbs of some of them are buckled by minor folds with their axial surfaces at high angles to the axial surface

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of the fold containing them. In Pembrokeshire these buckles are rarer and gentler than in South West England where many workers think that there has been some refolding.

4. The differences in structural style and history between the zones suggest that they evolved to some extent independently of each other. Thus because the zones are wholly or partially separated by major strike faults there is structural evidence to support the idea that some of the faults are significant boundaries.

ACKNOWLEDGEMENTS. I am grateful to Mr. T. R. Owen and Drs. S. C. Matthews and B. P. J. Williams for their helpful comments on Pembrokeshire geology and a draft of this paper. Professor D. L. Dineley, of Bristol University, is thanked for his continuing support of field work in Pembrokeshire. Alma Gregory drew the figure.

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THE STRUCTURE OF THE PERRANPORTH AREA, CORNWALL

by S. Henley

Abstract. It is suggested that the belt of steeply dipping rocks between Perranporth and Newquay originated in association with initial movement on the Perran Iron Lode after most or all of the f_2 folding, and that there is no large-scale inversion in the Perranporth area.

There have been two recent studies of the structural geology of the Perranporth area of the north coast of Cornwall, by Henley (1970) and by Sanderson (1971). The findings of Henley may be briefly summarised as follows : f_1 cleavage folds of tight to isoclinal style face north and have gently dipping axial planes. They are refolded by f_2 folds with northerly dipping axial planes. Within a belt 1 to 2 km wide, north of the northern boundary of the Gramscatho Beds, inter-limb angles of both f_1 and f_2 folds are smaller, and dips of cleavages and axial planes are close to vertical. The southern limit of this steep zone is defined by the Perran Iron Lode, a fault which appears to be of major significance (Henley and Fyson, in preparation).

Superimposed on these earlier structures are f_3 folds with horizontal axial planes, interpreted by Turner (1968) as due to vertical compression during granite emplacement.

The principal point of dissension with the later work by Sanderson concerns the nature and origin of the steep zone. Sanderson's observations, from study of only the Newquay-Perranporth coastal section, partly agree with results of the earlier study, and establish with some certainty the correlation of local f_2 events at Newquay and Perranporth. Some of his interpretations, however, are open to dispute. In describing a proposed structural cross-section from Newquay to Perranporth, Sanderson entirely disregards the Perran Iron Lode, and as a result is compelled to postulate that the steep belt is the middle

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limb of a monocline (which he correlates with f,) whose upper, antiformal hinge is rounded but whose lower, synformal hinge is angular (Fig. la). Such disparity of styles in the one major structure is not easily explained ; though Sanderson mentions that certain of the minor structures at Cotty's Point (SW757551) have similarly disparate styles, the Cotty's Point structure as a whole is complicated by the effects of f_3 folding which is more intense than further north, and folds of complex appearance might be expected from interference of the structural phases f_1, f_2 and f_3.

By inserting the Perran lode into the section (Fig. lb) one finds that no explanation of disparate styles is necessary.

Deformation associated with movement on the Perran lode also explains the observation (Henley 1970: 4.1, pl. 4.2) that f_2 folds as well as f_1 are flattened in a vertical east-west plane in the steep belt, a fact which Sanderson apparently ignores (though perhaps excusably in view of Turner's (1968) comment that f_1 closures are often very difficult to separate from tight f_2 closures). It seems, therefore, that the steep belt must have developed after much if not all of the f_2 folding.

Sanderson also asserts that the Perranporth structures are developed on the inverted limb of a major, north-facing, recumbent f_1 fold, but gives no evidence beyond dubious statements concerning vergence and facing directions. Way-up evidence in the coastal section is poor and inconclusive, quite insufficient for positive identification of facing directions. Furthermore, the presence of a large number of minor faults along the Perranporth cliffs gives a misleading impression of the vergence direction. The major fold structure at Cotty's Point is part of the inverted limb of a major north-verging f_1 fold rather than a south-verging cascade of minor f_1 folds in a regional scale inversion ; the scale of inversion at Cotty's Point is only of the order of tens or hundreds of metres. Conclusive way-up evidence is present near the northern margin of the Gramscatho Beds in inland exposures: at Ladock, for example, bottom structures, ripples, cross-bedding and graded bedding prove that the Gramscatho Beds are the right way up, and no evidence can be found for large scale inversion.

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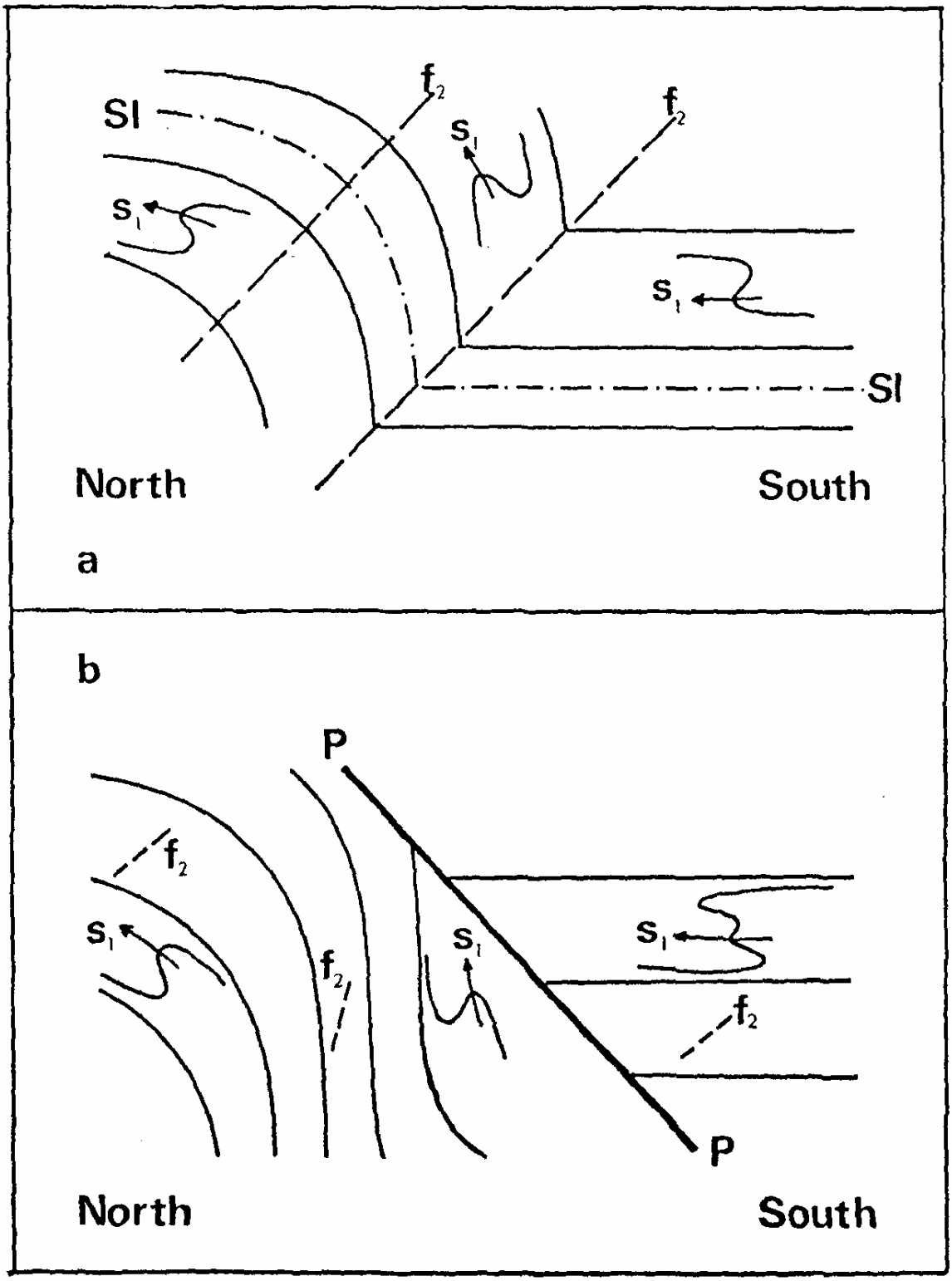

FIGURE 1. Comparison of structural interpretations of the Perranporth area, Cornwall. (a) after Sanderson (1971) ; steep zone resulting from f_2 folding, S1-S1, slide separating inverted and right-way-up limbs of a postulated f_3 recumbent fold. (b) suggested structure, based on inland as well as coastal evidence, taking into account the post-f_2 north-south compression and the probable effect of the Perran Iron Lode. P-P, Perran lode separating the steep zone caused by post-f_2 compression, from relatively undeformed rocks to the south.

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On the basis of the postulated inversion, Sanderson suggested that between Perranporth and Newquay there is an f_1 slide between the two limbs of a supposed recumbent fold (coinciding neatly with the position of the steep zone) which is either an f_2 (Sanderson) or post-f_2 (Henley) structure. He adduces Ripley's (1965) findings, that interlimb angles of f_1 folds decrease southward from Newquay, as supporting evidence. This decreasing interlimb angle, however, is seen also in the f_2 folds, and can be explained by a north-south *post-f_2* compression which culminated in reverse movement on the Perran Iron Lode (Henley and Fyson, in preparation). F_3 folding is sporadically present at the southern edge of the steep zone, and appears not to have been affected by these events ; formation of the steep zone and initial movement of the Perran lode appear, therefore, to post-date f_2 folding but pre-date f_3.

The normal displacement of the Perran lode, suggested by Henley (1971), is reconciled with the reverse displacement indicated above when it is realised that the fault has suffered a number of movement phases ; the nett throw is almost certainly in a reverse sense.

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CORRELATION OF FOLD PHASES IN S.W. ENGLAND

by David J. Sanderson

Abstract. Local developments of polyphase folding in Cornwall and north Devon are correlated. Three main divisions of fold phases are recognised and the features of the deformation outlined.

1. Introduction

Detailed studies of the polyphase nature of deformation in many parts of Cornwall and Devon have been established. In this paper a correlation of these local fold chronologies is suggested and the main phases summarised in a regional context.

Three main divisions in the fold histories are recognised (a) Primary deformation (associated with the development of slaty cleavage).

(b) Secondary deformation.

(c) Late phase deformation (syn- and post-granite structures).
Table 1 lists the local fold phases and their suggested correlation.

2. Primary Deformation

The primary deformation is characterised by the development of slaty cleavage and usually represents the most intense phase in the local deformation history. The primary deformation is itself polyphase, involving both the early F_1 folds of south Cornwall and the Newquay area, and the later F_1 folds of north Cornwall and north Devon (table I). At Polzeath, Roberts & Sanderson (1971) have demonstrated that the first phase folds in the south are refolded by the later primary folds in the north, which are hence F_2, in the deformation history. In south Cornwall the early primary folds face to the north or north-west, in contrast to the southerly facing later primary folds of north Cornwall. This polyphase primary folding produces the facing confrontation at Polzeath. In the extreme north of Cornwall the primary folds become upright and eventually overturned to the north in north Devon (see Sanderson & Dearman 1973 for a description of the attitude of primary folds).

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TABLE 1. *Fold correlations in Cornwall and north Devon*

	South Cornwall (Smith 1965, Stone 1966)	Newquay area (Sanderson 1971)	Polzeath area (Roberts and Sanderson 1971)	Tintagel-Boscastle (Dearman and Freshney 1966)	Bude-Hartland (Dearman 1967, Freshney and Taylor 1971)	North Devon
PRIMARY	F_1 (tight ———	F_1 close ———	F_1 ———→)			
			F_2 (←south facing	F_1 ———	F_1 upright ———	F_1 north facing→)
SECONDARY	F_2	F_2	F_3	F_2	F_2	king bands ?
LATE PHASE	F_3	F_3				
	F_4	F_4	F_4			kink bands

3. **Secondary Deformation**

The secondary deformation in S.W. England has produced a series of poorly correlatable local fold phases. In south Cornwall the F_2 folds generally have axial planes which dip steeply or moderately to the south, as at Porthleven (Stone 1966) and Godrevy Point (Smith 1965), or to the north, as at Perranporth (Sanderson 1971). These secondary minor structures appear to be related to major open folds or monoforms. In the Newquay area minor F_2 folds are related to the Watergate Bay antiform, now thought to be part of a monoformal complex (Sanderson 1971). In south Cornwall the F_2 folds may be related to the 'Truro antiform'.

Secondary folding is locally developed between Newquay and Tintagel, and includes the F_3 folds in the Polzeath area. Between Tintagel and Rusey beach F, drag folds are associated with low angle normal faults (Freshney 1965). In north Cornwall and mid Devon the secondary deformation is confined to some simple shear modification of F_1 folds on the flanks of the Culm synclinorium.

4. Late Phase Deformation

Two phases of widespread late folding are recognised in south Cornwall. Minor recumbent F_3 folds represent a phase of vertical flattening which Stone (1966) attributed to loading due to thrust or slide sheets. Lambert (1966) suggested that the F_3 folds (his F_2) were confined to areas around the granites. Detailed evidence on the nature and distribution of the F_3 folds by Turner (unpublished Ph.D. thesis, University of Newcastle upon Tyne, 1968) confirmed their relationship to batholith emplacement. F_3 folds are absent from the Bodmin and Dartmoor granites, however Dearman (1968) reports shear displacements along S_1 and S_2 cleavages in response to vertical F_3 flattening in the Lydford area.

The second phase of widespread late folding involves gentle or open upright folds (F_4) with sub-vertical crenulation cleavage which trends NNW-SSE. This phase is ubiquitous in south Cornwall and extends northward to the Polzeath area. These folds are unusual in that they represent ENE-WSW horizontal

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compression. Ghosh (1934) suggested that the granites consolidated under E-W compression with the formation of Q-joints and hence these may be related to F_1 folding. In north Devon the late phase kink bands with sub-vertical NNW-SSE kink planes may be a manifestation of this longitudinal compression.

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THE BRIDFORD THRUST

by E. B. Selwood and S. McCourt

Abstract. The gentle south dipping Bridford Thrust carries an Upper Palaeozoic sequence of Teign Valley facics, characterised by north facing folds, northwards over Crackington Formation (Upper Carboniferous) showing south facing folds of Meldon type. The significance of the structure is discussed.

1. Introduction

Recent work on structures at the southern margin of the Culm Synclinorium has been profoundly influenced by the pioneering studies of Dearman (1959) at Meldon. The structures and stratigraphy established in the British Railways Quarries have been confirmed by I.G.S. (Edmonds *et al.* 1968) and have been shown to persist along an E-W strike: traced southwards upright folds immediately north of Okehampton become increasingly over-turned towards the south, showing at Meldon axial planes dipping northwards at 45°. Dearman and Butcher (1959) further held that the progressive overturning of folds observed at Meldon continued down the west side of Dartmoor, developing into close recumbent folds and eventually merging with an extensive tract of recumbent isoclines. This change in fold attitude was related to the effects of granite emplacement. The current view expressed by Freshney and Taylor (1971) is that the transition "is due to the presence of a major overfold whose axis trends from the area of Wanson Mouth towards Okehampton," and that the zone of close recumbent folds "occupy the overturned portion of a major overfold caused by the overriding from the north of the upright folded blanket onto a more intensely folded infrastructure". An alternative view (Dearman 1969) involving recumbent folding of earlier upright folds has found little support to date.

To the east of Dartmoor a rather different picture has emerged. Although the broad structural pattern established on the Okehampton Sheet is continued along the strike onto the Exeter (325) Sheet (Simpson 1969), Chesher (1969) has shown that the Palaeozoic rocks flanking the granite on the Teignmouth (339) Sheet immediately to the south are not recumbently folded

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but thrown into a series of major folds which extend down the Middle Teign Valley towards Chudleigh. He notes that the folds are upright in the south but that they become progressively overturned *towards the north* when traced northwards, so that at the limit of the map, axial planes dip south at 78°. Inverted limbs are frequently cut and replaced by stretch thrusts.

The formations recognised by Chesher clearly continue northwards onto the Exeter Sheet, into an area bounded to the north and east by the River Teign. Little has been written on the geology of this area although Chesher, following I.G.S. practice, mapped a kilometer overlap onto the Exeter Sheet ; he recorded no significant structural change. Because of the fold style immediately to the north, an abrupt change in the facing of folds must take place hereabouts, and it was to determine the nature of this confrontation that the current investigation was undertaken. During 1972, as part of an undergraduate instruction program, Chesher's lines were continued northwards by mapping on a 6 inch scale. The results of this survey are presented here.

2. **Geology of the Bridford Area**

The major folds mapped in the Middle Teign Valley by Chesher are readily recognised because a steep (30-50°) east to north-east plunge off the granite exposes a number of distinctive horizons in a conformable sequence of Upper Devonian to Upper Carboniferous rocks. The full range of formations recognised by him cannot be mapped in the Bridford area. In part this is due to indifferent exposure, but mainly to the loss of character of shale horizons within the metamorphic aureole of the Dartmoor granite. On the accompanying map (Fig. 1) Hyner and Trusham Shales have been mapped as a single unit, and the Ashton Shale is undifferentiated from the Crackington Formation.

Immediately north of the Teignmouth Sheet Chesher maps a synclinal structure, the northern limb of which is cut by a late E-W wrench fault which displaces the granite and formational boundaries 15 m sinistrally. The fold structure is continued northwards quite concordantly, first into the Bridford anticline and then into the Bridford syncline. Both plunge 45° to 080 and show axial planes dipping 78° to the south. The structure, which is picked out particularly clearly in this area by dolerite

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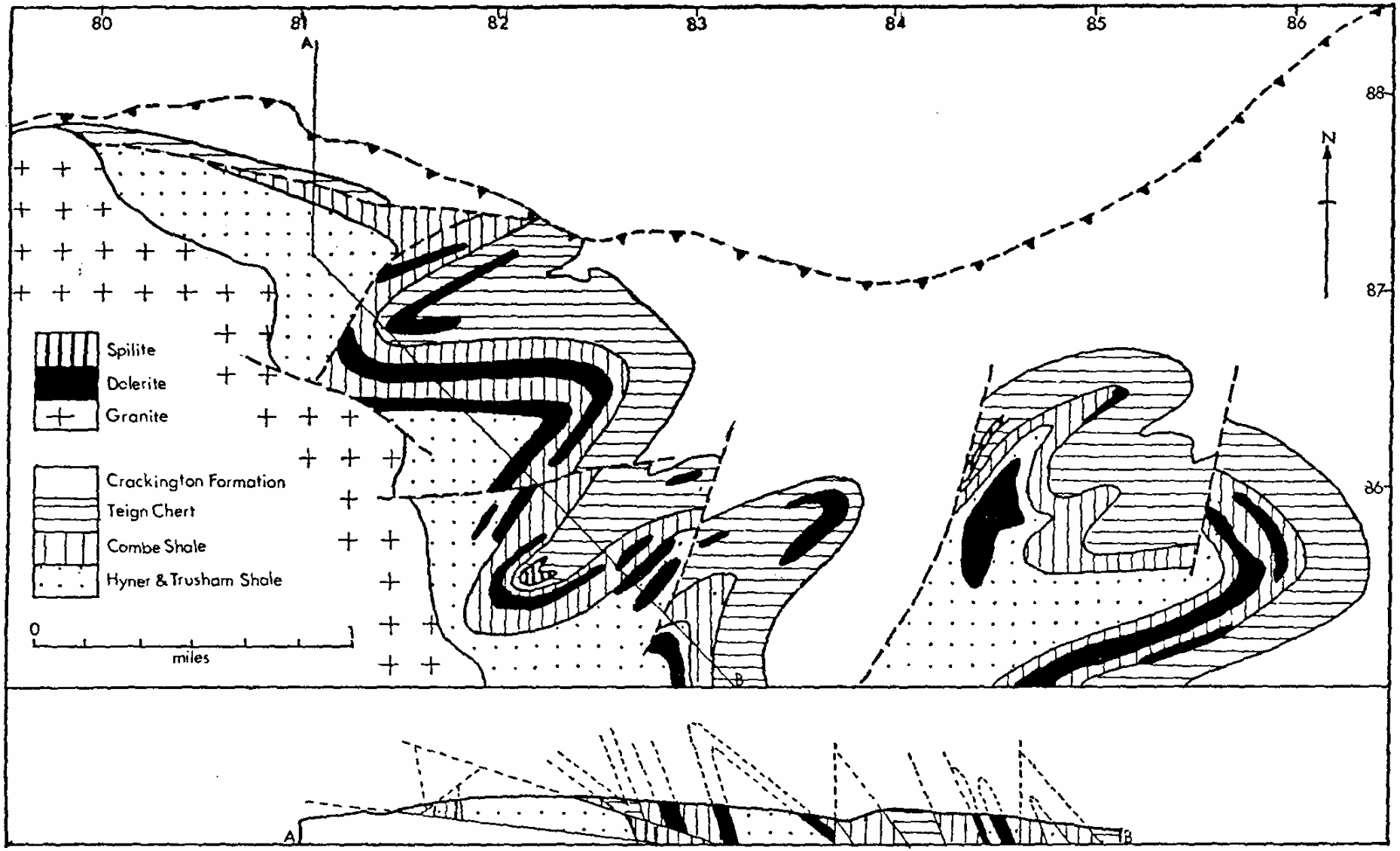

FIGURE 1. Geology of the northern part of the middle Teign valley.

Barbed line indicates Bridford Thrust. South and south-east part of map based on Chesher (1969).

sills and cherts in the Lower Carboniferous, ends abruptly north-east of Bridford against an important fault introducing Upper Carboniferous rocks. Adjacent to the fault, the rocks are severely brecciated and mineralised ; this is particularly evident at Birch Farm (SX 82208731) where the cherts run into the fault. A notable topographic break and the brecciation allow this fault to be traced north-westwards at a gentle inclination through the aureole to end against the granite 1.25 km south west of Steps Bridge. Outside the aureole, the trace of the fault plane V's gently down stream ; it crosses the River Teign near Sheldon, where a smash zone can be observed in the railway cutting, and eventually disappears eastwards beneath the New Red Sandstone deposits of Haldon at Idestone. An inclination of 7° to the south is indicated. This fault acts as a thrust, here named the Bridford Thrust, limiting and carrying the succession and structure of the Middle Teign Valley northwards over Crackington Formation, in which axial planes are characteristically inclined to the north.

Within the aureole the more varied lithologies of the Upper Devonian and Lower Carboniferous, which arc carried over the Crackington Formation, allow minor features associated with the thrust to be mapped, and it is evident that considerable disruption of strata has taken place. The upper limit of the dislocation is a thrust fault running north-eastwards from the granite margin at Rowdon Brook (SX 81058650) to join the main thrust. This minor thrust, which dips 17° to the south-east has carried the Bridford syncline northwards over an anticlinal structure. Between the two thrusts further disruption is evident: in particular, the northern overturned limb of the overridden anticline is cut by a steep north dipping fault. This fault, an accommodation structure disposed between the two thrusts, cuts out much of the lower part of the chert succession on the northern limb of the overridden anticline.

3. Interpretation

The confrontation of facing directions observed in the Teign Valley is thus explicable in terms of shortening associated with the Bridford Thrust. It is however difficult to ascribe great significance to the vergence of folds north and south of the Thrust, for no fundamental structural or stratigraphic difference can be identified across it. This is emphasized by Waters' report (pers. comm.) that the upright folds described by Chesher at the southern

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end of the Middle Teign Valley, when continued southwards become overturned to the south in the Lustleigh Fault Zone. This repeats the fold style north of the Bridford Thrust and it makes the structure south of the Thrust essentially anticlinorial. It thus appears that the folds between Exeter and the recumbent isoclines that occur to the south of a line from Chudleigh to Ilsington form part of a single structural and stratigraphic unit (see Simpson 1969) in which the facing of folds varies with the position on major anticlinorial and synclinorial folds. The Bridford Thrust has had the effect of bringing together two anticlinorial areas.

The observations reported here indicate that the Bridford Thrust pre-dates the intrusion of the granite and postdates not only the folding and associated stretch thrusting (Chesher 1969) seen in the Middle Teign Valley, but also the Meldon type folds to the north. The clear sense of northerly transport on the thrust is in perfect accord with the broad structural picture that has been developed on the Teignmouth Sheet (Waters 1970), where successive north facing and recumbently folded sheets have been moved northwards along gently south dipping thrust planes. At his northern limit of thrusting, Waters records recumbent isoclines thrust over the structures of the Middle Teign Valley.

It seems reasonable to suggest therefore that the Bridford Thrust was formed during the period of major overthusting from the south, and that the high level structures of the Teign Valley were in existence at least before the final phase of thrusting.

4. Significance

The Meldon fold, to date, has always demanded special explanation: thus, in the Okehampton Memoir, "The thrusting of a shelving 'basement' northwards beneath the margin of the geosyncline would produce overriding from the north". The view offered here is that the Meldon fold, far from being a unique structure, is an integral part of a fold pattern which, as developed on the east side of Dartmoor, is continued southwards at least into one further anticlinorial structure. This being so, the change from overturned to close recumbent folds observed west of Dartmoor need not be associated with underthrusting or any other special mechanism ; it need only signal the passage into a synclinorial fold zone.

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Nowhere does the Meldon-Teign Valley fold style pass gradationally southwards into recumbent isoclines ; where investigated in detail, the contact has proved to be faulted. On the coast and inland towards Launceston. it is marked by the Rusey Thrust along which southerly directed movements are recorded (Freshney *et al.* 1972), and east of Dartmoor by the Narrowcombe and Bickington Thrusts, which show a northerly sense of transport. The impasse of conflicting transport directions can be resolved if it is remembered that, although these thrusts separate the same structural zones, the movements have occurred at quite different times. Thus the movement on the Rusey Thrust is a much later tectonic event than the northerly directed thrusting. Alternatively, because the Rusey Thrust is acknowledged to have a long and complicated history, the observed throw may represent a re-activation with a reversed direction of throw of a more fundamental dislocation.

Between Launceston and Dartmoor the line of the Rusey Thrust has not been fixed precisely ; relations here seem particularly confused. However there is evidence, at present under investigation, which could suggest that successions of southern origin have been carried over Meldon type folds.

ACKNOWLEDGEMENTS The authors are particularly indebted to Dr. J. A. Chesher whose work in the Middle Teign Valley stimulated this study. The help given by Prof. S. Simpson, Dr. J. M. Thomas and students involved in the original mapping is gratefully acknowledged.

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AN INTERPRETATION OF NEW STRATIGRAPHIC EVIDENCE FROM SOUTH CORNWALL

by P. M. Sadler

Abstract. Detailed mapping and palaeontological investigations in the Roseland area of south Cornwall have enabled the compilation of two stratigraphic sequences. The first includes greywackes and limestones of Middle Devonian age. The second comprises a thin, Lower and lowest-Middle Devonian volcanic sequence, which probably rests directly on Ordovician quartzites and is succeeded by greywackes. A correlation between the greywackes seems reasonable but cannot yet be proven.

1. Introduction

The main contribution to the stratigraphic investigation of south Cornwall has been made by Hendriks. In two important papers (Hendriks 1931, 1937) she proposed a succession based on detailed lithological correlations. She indicated that most of the succession was probably Devonian and that Ordovician and Silurian rocks were of only local extent. These proposals confirmed the far-sighted conclusions of Sedgwick (1852). They rejected Reid's (1907) assumption that the Ordovician age, which had been long established for certain quartzite faunas (Peach 1841 ; Salter 1864), was of regional significance. Hendriks' conclusions were incorporated in a Geological Survey revision of the Lizard area (Flett 1946) but Reid's map ("New Series" sheet 353 - Mevagissey) remained unrevised. Further, the loosely defined "crush-breccia" structure which Reid described from large parts of the area (and toward which sheet 353 is very biased) seems to have prompted subsequent fault-breccia (Hendriks 1937) and "Wildflysch" (Dearman 1971) interpretations.

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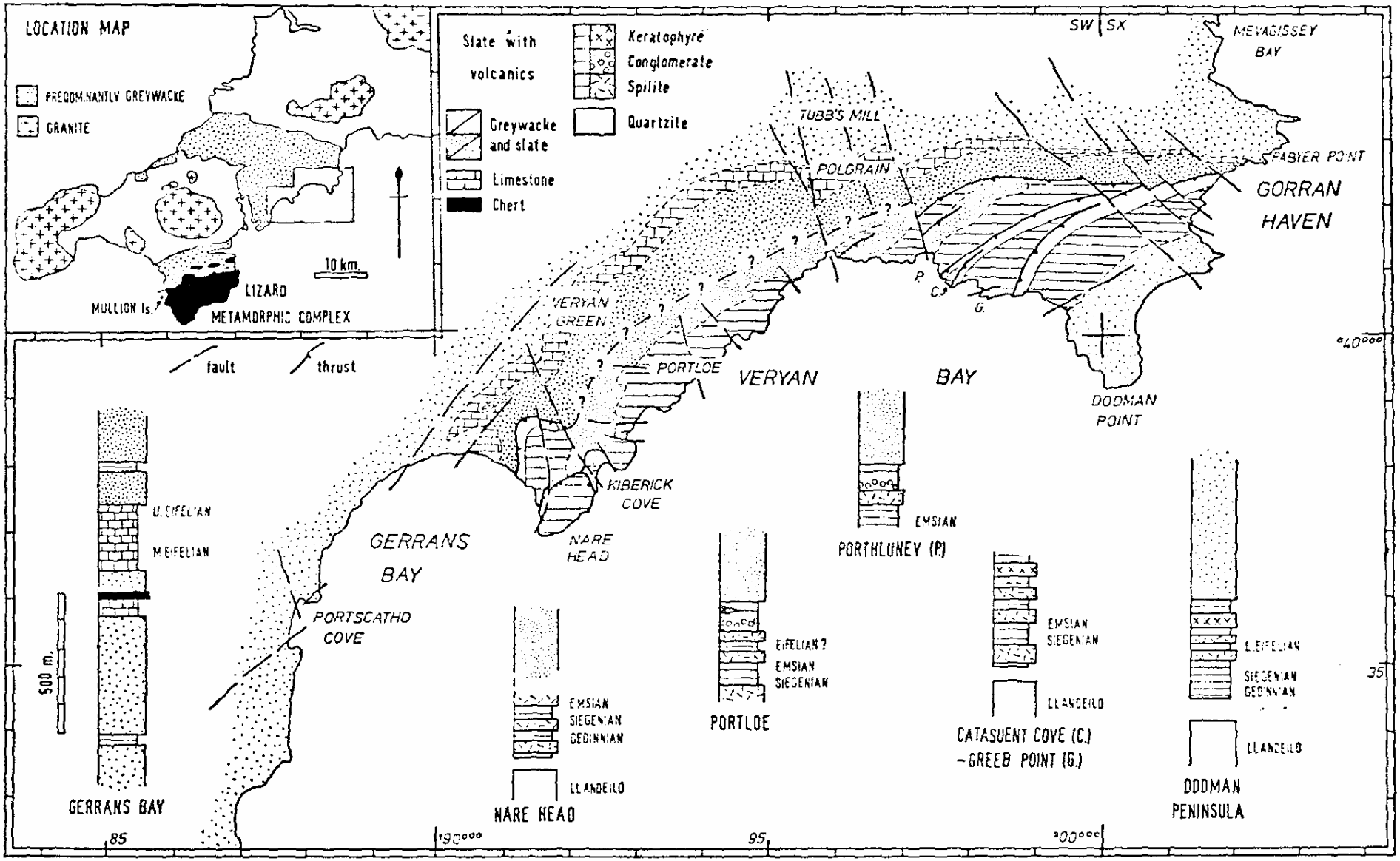

FIGURE 1. Generalized map and columnar sections showing the distribution of the main lithological units in the Roseland area of south Cornwall, with stratigraphic ages as indicated by conodont and trilobite faunas.

Recently Hendriks (1971 ; Hendriks, House and Rhodes 1971) has been able to cite a limited number of conodont determinations to prove the presence of Devonian rocks. No structural account was given, but it is significant that these two studies attempted to recognise stratigraphic sequences in sections previously interpreted as "crush-breccia". In the east Meneage district, where the concept of crush-breccia development was first formulated (Hill 1900), Lambert (1965) has identified a series of primary sedimentary lithologies.

The recent investigation (Sadler 1973) was an attempt to develop the leads given in these later publications by combining a thorough collection of conodont faunas with lithological and structural mapping. The principal area of study is shown in Figure 1. It was found there that evidence was readily available, in the form of minor folds in bedding, for the reconstruction of the structural events. Conspicuous, recurrent lithological sequences could then be built into a relatively simple stratigraphy and finally the determination of conodont faunas corroborated and refined this lithostratigraphy. This paper presents the palaeontological evidence for the proposed stratigraphic succession. The wider significance of these proposals and the structural synthesis will be treated elsewhere (Sadler, in preparation).

The area requires two distinct palaeontological approaches. In the north and west sides of Gerrans Bay, Hill (1898) and Hendriks (1937) have demonstrated the presence of a sequence of greywackes and slates (for which the name Gramscatho is retained) including a unit of thin, bedded limestones (here termed Veryan limestones). The task here is to date the conodont faunas developed through a recognisable sequence of limestones, which Rhodes (Hendriks *et al.* 1971) has already shown to be, at least in part, middle Devonian. In contrast, outcrops on the east side of Gerrans Bay, and through much of Veryan Bay and Gorran Haven comprise mostly volcanic rocks and conglomerates, set in black slates (here termed the Roseland volcanics). It is principally to these lithologies, among which limestones occur only as small lenses, that the "crush-breccia" and "Wildflysch" models have been applied. Consistent cleavage orientations tell against the first model. Secondly, bedding and primary sedimentary structures are so frequently observed, that the limestones must occur, not

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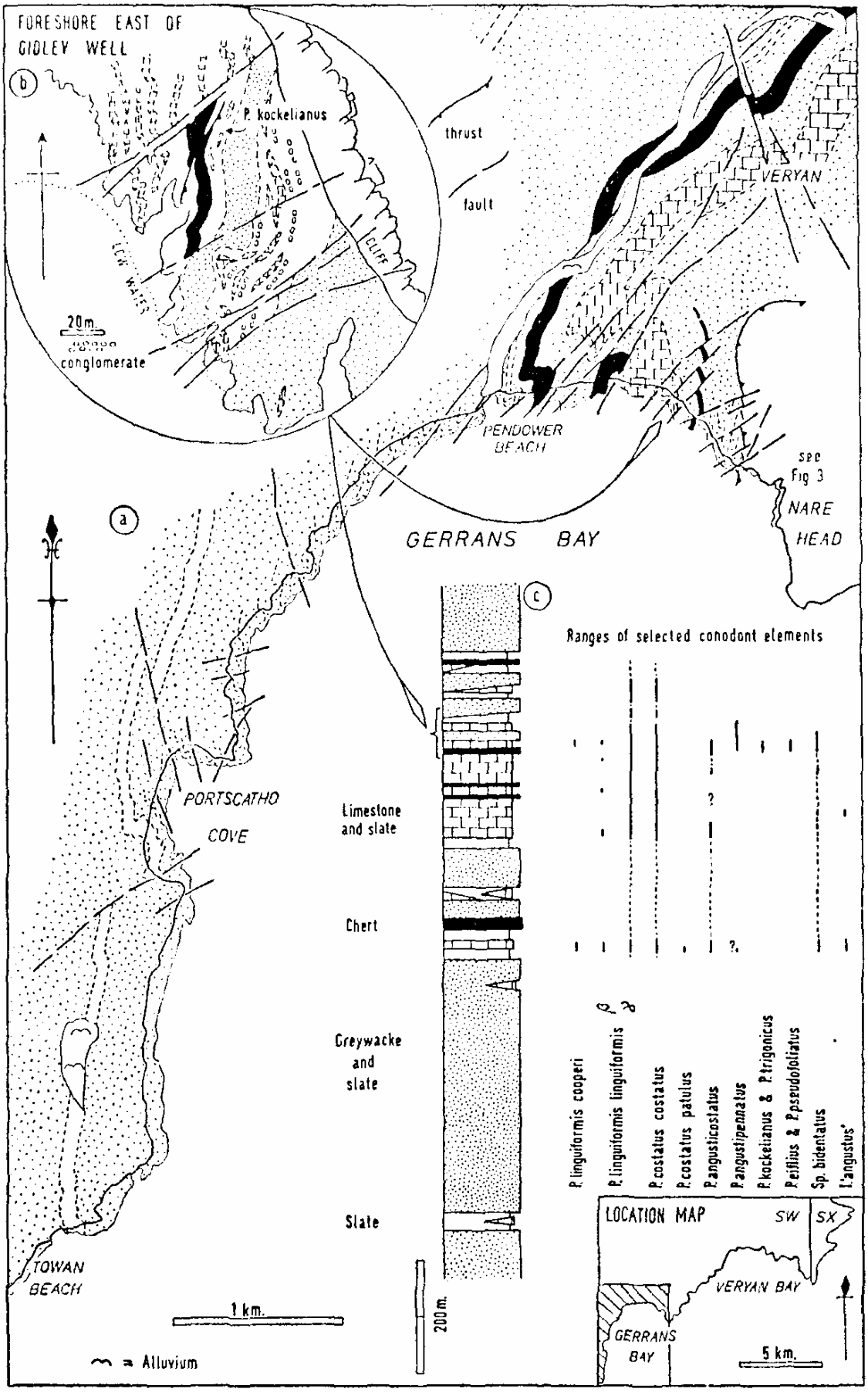

FIGURE 2. Map (a) and columnar section (c) of the lithological sequence in Gerrans Bay, with the ranges of stratigraphically important conodont elements (c) and enlarged detail of a critical foreshore outcrop east of Gidley Well (b).

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as clasts in a chaotic tectonic or sedimentary breccia, but as part of a normal folded sequence. The first question here is whether the conodont determinations support this contention. Consequently each fauna must be treated individually. No sequence of sampled horizons is assumed.

2. **The Veryan limestones**

These limestone beds outcrop on the foreshore of Gerrans Bay between Pendower Beach (SW 8984 3814) and Gidley Well (SW 9089 3814). Hill (1898) demonstrated their conformable position within a series of greywackes and slates. Mapping in Gerrans Bay (Fig. 2a, c) confirms this. The sequence corresponds to Hendriks' over-ridden region (1971) or Gramscatho succession (Hendriks *et al.* 1971). Hendriks (1937 ; in Holwill and House 1969) suggested the presence of an outlier of younger rocks near Gidley Well and advised that greywackes outcropping east of there might be a repetition of those seen below the Veryan limestones. A reinterpretation of the critical outcrop (Fig. 2b) recognises the importance of normal faulting and shows the essential continuity of the sequence.

52 conodont faunas (from 69 samples) are now available to demonstrate age and sequence in limestone units in the Gramscatho greywackes. *Polygnathus costatus costatus* Klapper (formerly referred to *P. webbi)* and *P. linguiformis linguiformis* Hinde (8 morphotype of Bultynck 1970) occur throughout the Gerrans Bay outcrop and in equivalent, thinner limestone units near Veryan Green (SW 9217 4031), Polgrain (SW 958 424) and Pabyer Point (SX 0215 4264).

A thin unit of limestone lenses towards the upper limit of the Veryan limestones near Gidley Well (Fig. 2b) is characterised by the appearance of two additional groups of conodont elements: *P. kockelianus* Bischoff and Ziegler - *P. trigonicus* Bischoff and Ziegler - *P. angustipennatus* Bischoff and Ziegler, and *P. eiflius* Bischoff and Ziegler - *P. pseudofoliatus* Wittekindt. This association is a clear indication of the upper Eifelian. Bischoff and Ziegler (1957: 129-175) and Wittekindt (1965: 627) recognise a *P. kockelianus* zone at the top of the Eifelian. It should be noted, however, that higher limestones in Gerrans Bay contain no specifically Givetian conodont elements. In the type Couvinian

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succession (Bultynck 1970) and in the Devonian of New York and Illinois (Klapper *et al.* 1971) P. *kockekianus* and P. *eiflius* disappear before the base of the Givetian. There is therefore no positive evidence that the Givctian is present in Gerrans Bay.

P. costatus costatus appears in the mid-Eifelian of New York, Illinois and Indiana (Klapper *et al.* 1971) and in the middle of the type Couvinian (Bultynck 1970). The dominance of the u morphotype of P. *linguiformis linguiformis* also indicates that the main part of the. Veryan limestones unit 'is mid-Eifelian *(Spathognathodus bidentatus* zone of Bischoff and Ziegler 1957). The association of P. *costatus costatus* with *P. costutus patulus* Klapper in the lowest Veryan limestones suggests a low-mid-Eifelian age (Klapper 1971). Limestone lenses from Portscatho Cove yielded only non-diagnostic conodont elements.

3. The Roseland volcanics

The sequence described from Gerrans Bay is interrupted near Carne (SW 9126 3777) by a thrust fault which introduces an orthoquartzite unit. This quartzite has yielded the trilobite *Crozonaspis* Henry (Sadler, in press) - a genus characteristic of the Llandeilo of the Armorican Massif, Spain and Morocco.

The succeeding volcanic sequence must be reconstructed from the zone of imbricate thrust faults between Pennarin Point and Nare Head (Fig. 3). These outcrops present complete and partial sections of a characteristic lithological sequence - a volcanic interval comprising lava or lava-breccia and coarse tuff (mostly spilitic but locally keratophyric) is succeeded by black slates. The lower part of the slate interval includes fine green tuff beds and some coarse agglomerate ; the upper part is characterised by small graded lenses of coarse green volcanic debris. Between these two slaty facies there occurs a coarse polymictic conglomerate succeeded by impersistent beds of quartzite and lenses of coarse crinoidal or orthocone limestone. Differences between the various lava intervals and the subsequent determination of conodont faunas from the limestones indicates that at least two of these lithological sequences are present. The associated quartzites yielded no fossils. It is fundamental to this study that only those quartzites containing the characteristic trilobite-brachiopod fauna are attributed an Ordovician age.

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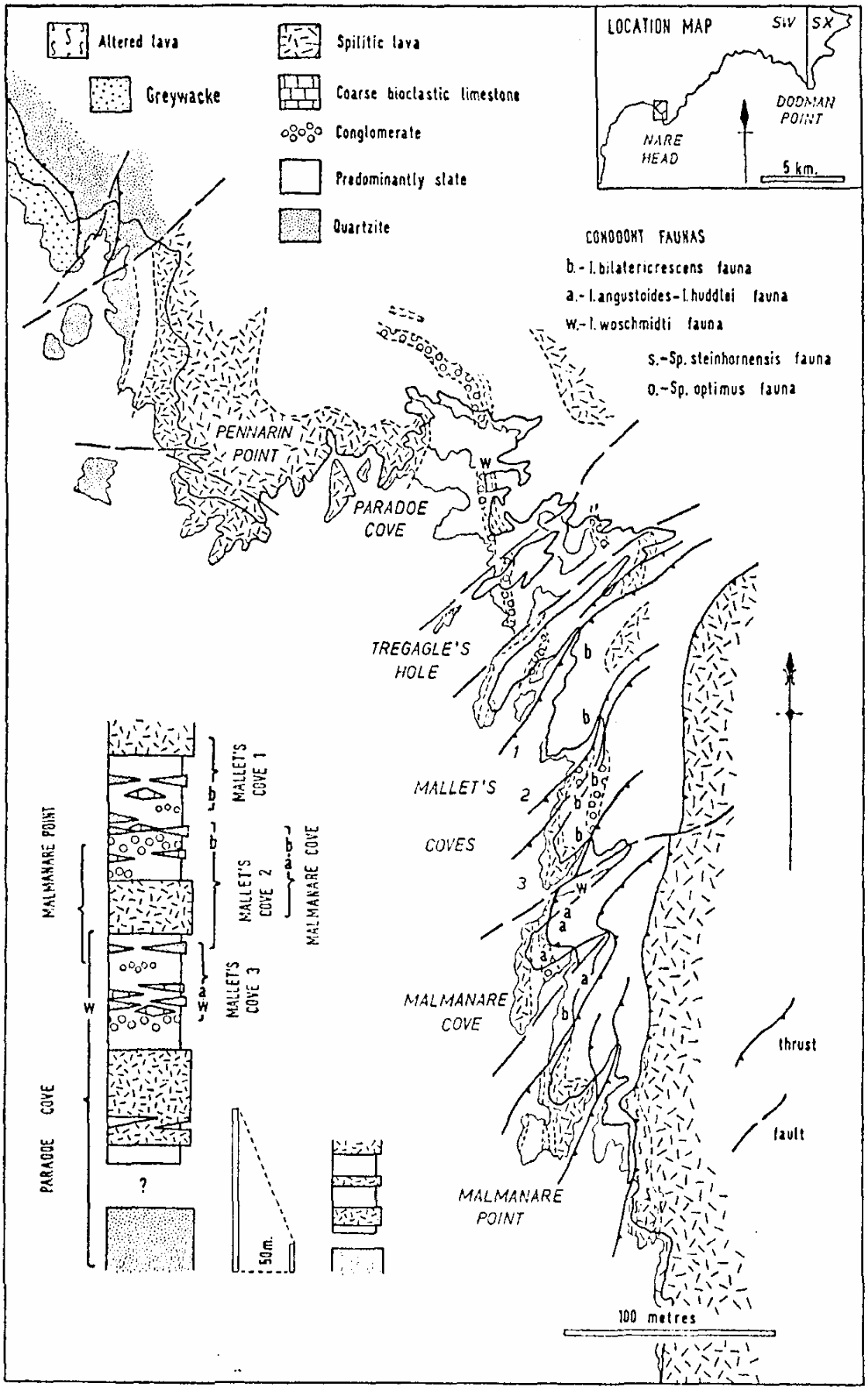

FIGURE 3. Map and columnar reconstruction of the stratigraphic sequence from the imbricate thrust zone on the west side of Nare Head. Localities of diagnostic conodont faunas indicated. (Small scale column drawn for comparison with Fig. 2c.).

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A limestone tens from tile sequence seen in Paradoe Cove (SW 9146 3765) has yielded a conodont fauna including *Icriodus worschmidti postwoschmidti* Mashkova (sensu Ziegler 1971:235) and *I.* sp. cf. *I. rectangularis s.l.* (s.l. Ziegler 1971). These icriodiform elements have been reported from north-east Spain (Carls and Gandl 1969) and from Nevada and the Yukon (Klapper *et al.* 1971) where associated shelly faunas and graptolites indicate a Gedinnian age. *I. woschmidti* was also recovered from a limestone on the south side of Mallet's Cove 3 (SW 9151 3748). Two limestone lenses immediately above this yielded conodont faunas including *I. huddlei curvicauda* Carls and Gandl and three "subspecies" or *I. angustoides* Carls and Gandl. This icriodiform association has been described from Spain (Carls and Gandl 1969 ; Carls 1969) where associated brachiopods indicate a lower Siegenian age. From this same locality in Mallet's Cove 3, Rhodes (cited in Hendriks 1971 : 120) reported an element resembling *Spathognathodus transitans* Bischoff and Sanneman - indicative of high Gedinnian (Ziegler 1971, chart I). Associated with the demonstrably Siegenian limestones, a lens was found containing numerous orthocones and a pelecypod (compare Hendriks 1971 : 120). It failed to yield diagnostic conodonts.

A limestone higher in the same sequence, on the south side of Malmanare Cove (SW 9150 3744), again yielded *I. h, curvicauda* and *l. angustoides*. At a lower structural level in Malmanare Cove further limestones occur in the same characteristic lithological sequence and have yielded two diagnostic faunas. From the north side of the Cove (SW 9149 3745) *I. h. curvicauda* and *1. angustoides* were again recovered. Above this, on the south side of the cove, *I. bilatericrescens bilatericrescens* Ziegler and *I. bilatericrescens* ? *beckmanni* Ziegler were found. With a résumé of the associated macrofossils in Spain and Germany, Ziegler (1971: 244) has demonstrated the mid-Emsian age of these conodont elements. Limestones in an analogous position on the south side of Mallet's Cove 2 (SW 9150 3753) have yielded the same elements plus *I. huddlei celtibericus* Carls and Gandl and *Panderodus striatus aratus* Carls and Gandl - all typical of the mid-Emsian in Spain.

Conodont elements from the same association have also been recovered from limestones on the north side of Mallet's Cove 3 (SW 9151 3751), on the north side of Mallet's Cove 2 (SW 9150 3756), and on the south side of Mallet's Cove 1(SW 9150 3759).

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Most of these limestones also yielded internal moulds of gastropods and ostracods.

A synthesis of this information from the imbricate zone is suggested in the columnar section in Figure 3.

Lying on a low-angle thrust above this imbricated zone are the well-formed pillow lavas and coarse syenitic core of Nare Head itself. Bleached laminated limestone lenses among the pillows near Rosencliff (SW 9214 3732) produced no fossils. From coarse crinoidal limestones north of the syenitic core (SW 9240 3758) only gastropod moulds were recovered. These lavas cannot simply be dated by comparison with the perplexing, isolated occurrence of Frasnian (Hendriks *et al.* 1971) pillow lavas on Mullion Island (Fig. 1, inset). Lava intervals in the Roseland volcanics can now be shown to include Lower and lower-Middle Devonian flows. They are lithologically variable and are locally swollen by thick pillow lavas (e.g. Great Perhaver Point, SX 015 420 and Long Point, SW 699 411). Coarse grained facies are encountered again on the Blouth (Fig. 4c). A comparable thick pillow lava also occurs at Tubb's Mill (Fig. 1). It faces upward and appears to dip beneath the Veryan limestones.

The basal thrust at Nare Head appears to be synformal and outcrops again on the west side of Kiberick Cove (Fig. 4). Here it over-rides greywacke and slate which may succeed the Roseland volcanics beneath Nare Head. That this is the case can be demonstrated in a major inverted limb outcropping along the west coast of Veryan Bay (Fig. 4). The succession reconstructed from Nare Head is recognisable at Manare Head. As it is traced northward and upward (stratigraphically), the lava intervals become thinner and more numerous. The limestones and quartzites disappear. A thick, very coarse conglomerate outcrops at May's Rock. Above this a thin lens of keratophyric material is the last trace of volcanic activity. Next, slates with graded conglomerate layers pass upward into greywackes and slates, indistinguishable from those in Gerrans Bay. Figure 4a also shows how south of Manare Head the older parts of this succession are thrust over progressively younger parts.

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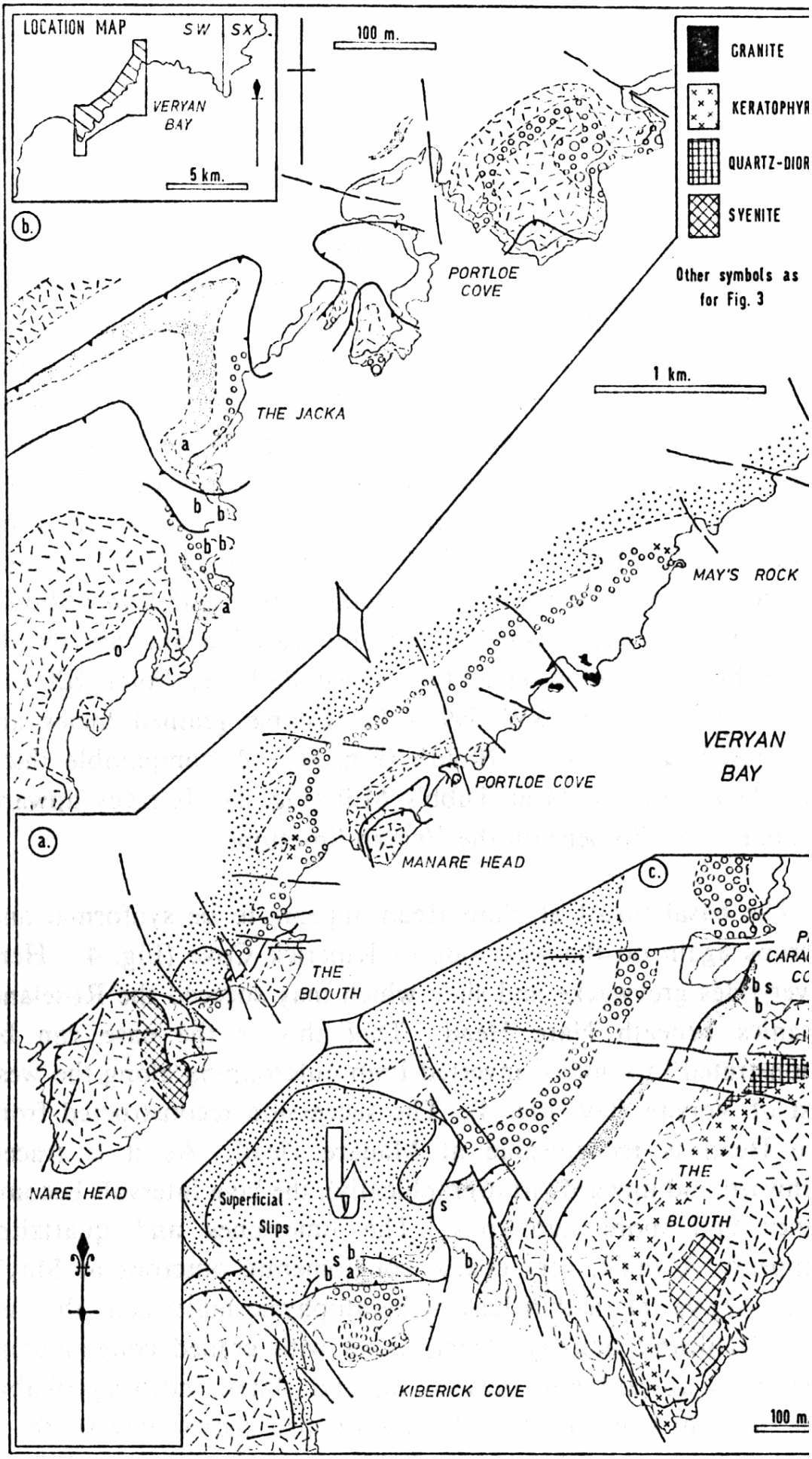

FIGURE 4. Generalized geological map of the west coast of Veryan Ba with details of the location of diagnostic conodont faunas the Jacka (b) and the Blouth (c).

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I. woschmidti faunas have not been recovered from this section. An orthocone limestone immediately above (stratigraphically) the lava interval on Manare Head (SW 9361 3895) yielded *Sp. optimus* Moskalenko. The type material of *Sp. Optimus* and examples from N. America are Siegenian (Klappcr 1969). A limestone occurring somewhat higher, above a quartzite interval (Fig. 4b), has yielded *I. h. curvicauda.* This icrio;liform element is found without *I. angustoides* in the highest Siegcnian of north-east Spain (Carls and Gandl 1969). Above this the characteristic conglomerate is found and then several limestone lenses yielding the conodont fauna previously shown to be mid-Emsian - *I. b. bilatericrescens, I. b. multicostatus, I. b? beckmanni, I. h. huddlei* Klanper and Ziegler and *I. h. celtibericus.* The highest of these also includes *P. lenzi* Klapper and. *I. sigmoidalis* Carls and Gandl. An associated orthocone limestone produced no conodonts.

At the south end of the Jacka (SW 9365 3914) the sequence is repeated in a lower thrust slice. Here a limestone immediately above the quartzite bed produced only *I. h. celtibericus.* Carls and Gandl (1969) record that this is the only icriodiform found in the lowest Emsian between the *I. h. curvicauda* and the *I. bilaticrescens* faunas. Higher limestones occur in this thrust sheet at the northern end of the Jacka and in Portloe Cove but (lid not yield diagnostic conodonts.

The younger parts of the sequence to the north contain very few limestones. However, from the northeast of Portioe Cove Rhodes (cited in Hendriks 1971 : 121) obtained a conoelont fauna which is probably Eifelian.

The full *I. bilaticrescens* association occurs again in limestones in Parc Caragloose Cove (SW 9290 3838 : Fig. 4c) where an associated orthocone limestone produced *Sp. steinhornensis steinhornensis* Ziegler - an equivalent mid-Emsian element (Ziegler 1971). The same sequence is found again in Kiberich Cove but here the outcrop is complicated by an extensive, two-stage rotational slump (Fig. 4c). The *Sp. st. steinhornensis* fauna shown in the north-east corner of the cove is from an orthocone limestone made available by a recent cliff fall.

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The transition from volcanic facies to greywacke outcrops again, twice, in the rear of Veryan Bay (Fig. I). In the volcanic sequence at Battery Point (SW 9717 4108) the mid-Emsian *Sp. st. steirnhonnensis* was recovered from two lenses of orthocone limestone. In a faulted repetition of the same horizon in Porthluney Cove (SW 9757 4104) the orthocone lenses were unproductive but from a slightly lower crinoidal limestone *I. h. celtibericus* was found.

As Figure 1 shows, the structure on the west side of Veryan Bay is relatively simple, comprising three overthrust slices. The Ordovician age of the basal quartzite can now be proven for each overthrust sheet by the occurrence of Llandeilian trilobites including *Neseuretus* (*N*) *tristani* (Brongniart), *Kloucekia* (*K.*) *mimus* and *Bathycheilus sp.* aff. *B. perplexus* (Barrande) (Sadler, in press).

Very little is available from the lowest thrust sheet but limestone lenses in Catasuent Cove (SW 9765 4083) yielded the *I. h. curvicauda - I. angustoides* association, succeeded by *I. h. curvicauda* alone. Comparison with Spanish occurrences (Carls and Gandl 1969 ; Carls 1969) suggests lower and upper Siegenian ages respectively. Hendriks *et al.* (1971) have discussed the significance of the *Scyphocrinites* found at the same locality (Bather 1907). The matrix from this specimen failed to yield diagnostic conodonts (S.C. Matthews, pers. comm.). A short distance east of Catasuent Cove (SW 9766 4080), a higher limestone horizon produced three mid-Emsian *I. bilatericrescens -I. huddlei* associations.

The second overthrust sequence has yielded no diagnostic conodont forms. Toward the top of the available outcrop a thin greywacke unit interrupts the volcanic sequence. The highest unit seen is a thick keratophyric lava or tuff at Greeb Point (Fig. 1). This keratophyre occurs again in the highest thrust sheet shortly below the base of the greywackes. This highest structural level includes a complete outcrop through the volcanic facies which here comprises distinct massive and pillowed lavas with relatively less conglomerate. Unfortunately, because bedding is here almost vertical, the massive quartzites have slumped at the coast, obscuring the lower parts of the succession. Nevertheless, at Great Perhaver

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Beach (SX 0161 4229) a Gedinnian conodont fauna - *I. woschmidti postwoschmidti, I.* sp. cf. *I. rectagularis* s.l., *Pelekysgnathus serratus elongatus* Carls and Gandl and *Pe. serratus* sbsp. A Carls - has been obtained from a small, solid outcrop of limestone accessible between the fallen masses of quartzite. Further, a fallen block including limestones has yielded *I. h. curvicauda.*

The sequence in this third overthrust mass exposed in Veryan Bay is extremely similar to that available in Gorran Haven. An horizon of coarse limestone lenses in the Veryan Bay outcrop (SW 9884 4061) of the volcanic facies his yielded *P. costatus* sbsp. indet. and *P. linguiformis linguiformis* (α morphotype, Bultynck 1970). The latter morphotype is found toward the base of the type Couvinian and type Eifelian sections (Bultynck 1970). This fauna confirms the implication from Rhodes' conodonts from Portloe Cove that the Roseland volcanics unit extends into the middle Devonian.

Taken together all these conodont faunas indicate the existence in south Cornwall of a condensed volcanic sequence spanning, in its lower part, almost all of the lower Devonian and extending into the Eifelian. There is one indication that the conodont sequence may not be complete: the rather problematic *I. pesavis* - *Ancyrodelloides trigonica* fauna (Ziegler 1971) has not been found. The yield of diagnostic faunas was not high (35 diagnostic and 20 non-diagnostic faunas from 151 samples) but their age determinations certainly add weight to the contention that the limestones are not part of a tectonic or sedimentary melange. No evidence of the Silurian or higher Ordovician was obtained. Orthocone limestones similar to those previously referred to the Silurian (Reid 1907 ; Hendriks 1937) are found to be relatively common and to yield Siegenian and Emsian conodonts. By analogy with similar successions in Normandy and Brittany, Figures l and 3 suggest the presence of a non-angular unconformity above the Llandeilo quartzite unit.

4. Discussion

The stratigraphy of south Cornwall is quite distinct from that north of the outcrop of the Gramscatho greywackes. The interpretation proposed here indicates several Armorican affinities in the older part of that stratigraphy, but the thick, apparently

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Middle Devonian Gramscatho greywackes remain peculiar to south Cornwall. To locate this period of greywacke sedimentation in time, one can at present only try to produce brackets of ages from adjacent lithologies. Unfortunately these brackets are often loose and still cannot be closed.

Figure I indicates four greywacke bodies which might deserve separate consideration. In Gerrans Ray two greywacke phases are identified : one finishing in the early Eifelian, the other commencing near the close of the Eifelian. Elsewhere in the Gramscatho Outcrop (Fig. I inset) the break may not be identifiable. The Outside limits of the greywacke deposition cannot be properly fixed. Hill (1898) claimed that the greywackes pass into the slates of the Falmouth and Mylor Series. However, it is still true as Hill noted in 1900, that even the relative age of these beds is uncertain. Edmonds *et al.* (1969: 29) and Sanderson (1971) have questioned the stratigraphic and structural simplicity of the northern limit of' the Gramscatho ereywackes where they are in contact with the lower Devonian Meadfoot Beds.

Lithological similarities between the sections drawn up from Veryan Bay indicate that all tile greywacke seen there belongs to one unit. The greywackes of the Dodman peninsula are Reid's (1907) "Dodman Phyllites". This unit is not lithologically distinct from other greywackes in Veryan Bay (see also McKeown 1962). Its position in the lithostratigraphy is exactly analogous and its northern boundary is transitional, not a fault. The base of the Veryan Bay greywackes is lower Eifelian, or younger. This qualification is important because the underlying volcanic sequence is shown to be in part condensed. If the abundant limestone below the keratophyre at Greeb Point could be dated this end of one bracket would be considerably tightened. Large volumes of this rock have been dissolved but only sponge spicules have been recovered.

The base of the greywackes in Veryan Bay cannot yet be located with respect to the Gerrans Bay Succession. Although the greywacke facies seem to be identical the proper course is to recognise two separate stratigraphical sequences and to acknowledge that they cannot yet be shown to be contemporaneous. (One consequence of this is that the nature of the boundary between the two greywacke units which outcrops west of Portloe

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must remain uncertain). It is still possible that the base of the greywackes mapped in Veryan Bay corresponds to the base of the Gerrans Bay greywackes, but there are also tentative indications that greywacke deposition started later in the former case.

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THE CAMEL ESTUARY PLEISTOCENE SECTION WEST OF TREGUNNA HOUSE

by B. B. Clarke

Abstract. This section consists of two solifluction earths, separated by a coarse boulder gravel, suggested to be a till which originated as a lateral moraine.

Boulders on the modern beach, derived from a gravel on the south bank of the Camel estuary 0.5 km west of Tregunna House, were marked on the One Inch Geological Survey map as erratics. In the accompanying memoir, Reid (1910) described the gravel as either ice rafted or a valley gravel derived from Bodmin Moor. Apart from brief references (Clarke 1961, 1969) these gravels have not been described since. Their close relation to those of Trebetherick, and the presence of Wolstonian ice on the continental shelf west of Pentire Head now makes this desirable.

At Tregunna House two solifluction earths are seen together. The lower, 1.5m thick, is coarse textured and slatey. It rests on the raised beach platform which is frost shattered. The upper, lm thick, is much finer, with matrix of clay, quite persistent and, where the lower head is missing, rests directly on the rock platform. 0.5 km west of Tregunna House the two heads are separated by a very coarse boulder gravel, 0.25 to 0.5m thick, which persists unbroken for 140m westwards. The stones are all partially rounded, most have at least one rather flat face, and many have three giving the appearance of dreikanter. Many stones have a very deeply pitted surface, suggesting frost weathering. A few smaller stones have a really smooth surface, suggesting wind polishing, but pebbles which are smooth ovoid, and clearly well water-worn are rare. The rocks are of great variety, compared with the present beach gravel, which is entirely slate. Most are igneous - granites, lavas, tuffs, fine ashes and greenstones. Sedimentary rocks are rare but one conglomerate with siliceous pebbles set in fine ash may be Gravel Caverns conglomerate. The larger stones rarely touch, and there is a complete absence of grading and bedding. Most boulders are at least 30 cm across and the largest are invariably vein quartz. When the rock boulders are broken open most show a selvedge of highly weathered material about 2 mm width. The matrix consists of small pebbles of the

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same rocks as the boulders, together with sand, silt and rock flour. This tine material is also ungraded and unstratified. The fine material has many splintered mineral fragments and rounded grains are rare. None of the rocks can with certainty be recognised as erratics.

The absence of shell and wood fragments, and the surface pitting suggest the gravel is of glacial rather than interglacial origin. Furthermore the absence of grading and bedding, the variety of rocks, and the highly variable condition in which they occur all suggest a till rather than a fluvio-glacial gravel. The position on the valley side in a constricted part of the estuary suggests a surviving fragment of lateral moraine. Elsewhere it has been shown possible for glacier ice to over-ride unconsolidated material such as the Lower Head, without destroying it, provided it has been saturated with water and then frozen solid. Thus in the Gower, Irish Sea till overlies unconsolidated raised beach, and at Tralee Bay till has been shown to overlie Older Head (Mitchell 1970).

Post depositional features are abundant and affect both the Boulder Gravel and the Lower Head. They include involutions, frost wedge pseudo-morphs, faults, cracks, clay injection, frost heaving and the upending of rock fragments to vertical. The same feature may affect both the Boulder Gravel and the Older Head. This is especially true of the faults. This feature is believed to be most common in head of Wolstonian age (Shotton 1965). The Younger Head is unaffected.

The direction of ice movement cannot be determined from the section. It could have been that a tongue of the Wolstonian ice-sheet from the channel (Mitchell 1967, 1972) entered the open mouth of the north facing estuary, deposited the Trebetherick till, and moved inland into the lower part of the Camel valley at least as far as Tregunna. Such intruding ice would gather a variety of stones from the river bed (at that time deeply incised owing to the low sea-level) and frost weathered and wind faceted material from the valley sides. Alternatively but less likely, the ice could have taken the form of a valley glacier originating in Bodmin Moor which moved down the Camel valley to join the sheet ice in the channel. However, there are no obvious glacial depositional features in the Camel valley, such as characterise

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the ground formerly occupied by Weichselian ice in the Wye valley, or the valleys behind Bantry Bay.

The main problem in interpreting this section, however, is one of time. The rock platform was clearly cut before the Older Head was laid upon it. Periods of cold are required for the frost shattering of the rock platform, the formation of the materials of the two solilluction earths and the till. Warmer periods are required for the saturation of the Older Head, the re-activation of the hard frozen ground to impose the peri-glacial structural features affecting both the Older Head and the till, and the erosion of these two deposits. If the Wolstonian age of the channel ice is accepted, the evidence here suggests this cold stage was to some extent complex, with milder periods at intervals of considerable duration. One such is required for the downslope solifluction of the Older Head following the cold period of initial frost shattering, but before the peak of the cold which brought the ice and the till. The surface of the Older Head and the tendency of' the boulders at the base of the gravel to be wrapped round with head suggests a second period of soliiluction towards the end of the Wolstonian cold stage.

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LOESS IN DEVON

by T. R. Harrod, J. A. Call and A. H. Weir

Abstract. Thin silty drift deposits occur at several upland sites in south and east Devon. The particle size distribution and mineralogy of these drifts suggests they are composed mainly of loess, which has been weathered and partly mixed with deposits beneath. A distant source of the Windblown silt is proposed.

1. Introduction

Detailed petrographic studies of soil profiles in England often show that the parent material is not uniform with depth, but is a layered sequence possibly involving two or more drift deposits as well as solid formations. Thus differences between horizons, even in the highest metre or so of a soil profile, may reflect lithological changes in a sequence of Quaternary deposits as well as changes caused by weathering and other pedological processes.

One of the most widespread drifts affecting soils in southern England is a silty deposit, which, on account of its blanket-like distribution and mineralogical uniformity, has been attributed by many workers to a thin sheet of loess that has lost many of its distinguishing characteristics by weathering, local reworking and admixture with other deposits. Relatively thick loess, with many of the important characteristics listed by Russell (1944) and others, occupies a few small areas in Durham (Trechmann 1920), east Kent (Pitcher *et al.* 1954) and other parts of eastern England. However, thin weathered and reworked silty drifts are now known to occur extensively on the Chilterns (Avery 1964 ; Avery *et al.* 1959, 1969, 1972), West Sussex Downs (Hodgson *et al.* 1967) and Coastal Plain (Hodgson, 1967), Derbyshire Limestone Plateau (Pigott 1962, Johnson 1971), Mendip Hills (Findlay 1965) and Lizard Peninsula (Coombe *et al.* 1956). The head and river brickearths of Kent, the Thames Valley and southern parts of East Anglia have a similar composition and origin, at least in part, and the Norfolk coverloam is also thought to be windblown silt derived from Weichselian glacial debris (Catt *et al.* 1971). The thin silty drifts are important as soil forming materials, because they are more easily worked and have larger available water

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capacities than most of the deposits beneath, which are mainly heavy clays, gravels, or hard rocks.

We describe here the distribution of thin silty drifts in Devon, found while surveying soils in many parts of the county, and report the particle size distribution and mineraluoical composition of their coarse silt fractions. The analytical results allow comparison of the drifts with the loess of eastern England.

2. **Distribution and Field Characlerislics**

Although a systematic survey of the soils has been completed only in selected parts of Devon (Fig. 1), almost all the county has been examined in lesser detail. This examination showed that silty drifts occur in three main types of situation :

A On the East Devon Plateau (180-275m O.D.) almost all the ground is mantled by thin (<50cm) silty drift. This overlies the Clay-with-flints-and-cherts of Ussher (1906), which corresponds to the Plateau Drift of the Chilterns, as defined by Loveday (1962). Locally the junction between the two is involuted. Harrod (1971) mapped Batcombe, Dunkeswell and Blackdown soil series in these areas, and Hook series in a few small areas where the silty drift is > 80cm thick.

B On the Devonian Limestone Plateau silty drift up to 1.5m thick and mapped as Ipplepen series (Clayden 1971 : 130) floors narrow, often steep-sided, dry valleys. Thinner layers also occur sporadically on interfluves, to give soils resembling the Nordrach series, and involuted pockets of silty material occasionally occur in the clayey substratum of such soils. The deposit is lower here than in any other parts of the county, being below 60m O.D. in places.

C Silty drifts of variable thickness occur sporadically on the higher ground fortned by the Budleigh Salterton Pebble Beds in east Devon, the Upper Greensand and Haldon Gravels of the Haldon Hills, and the Dartmoor Granite. They occupy <20% of the landscape, but with no obvious relation to physiographic features. On the Haldon Hills podzols developed in thin silty drift over flinty gravel were grouped with the Southampton series (Clayden 1971: 143).

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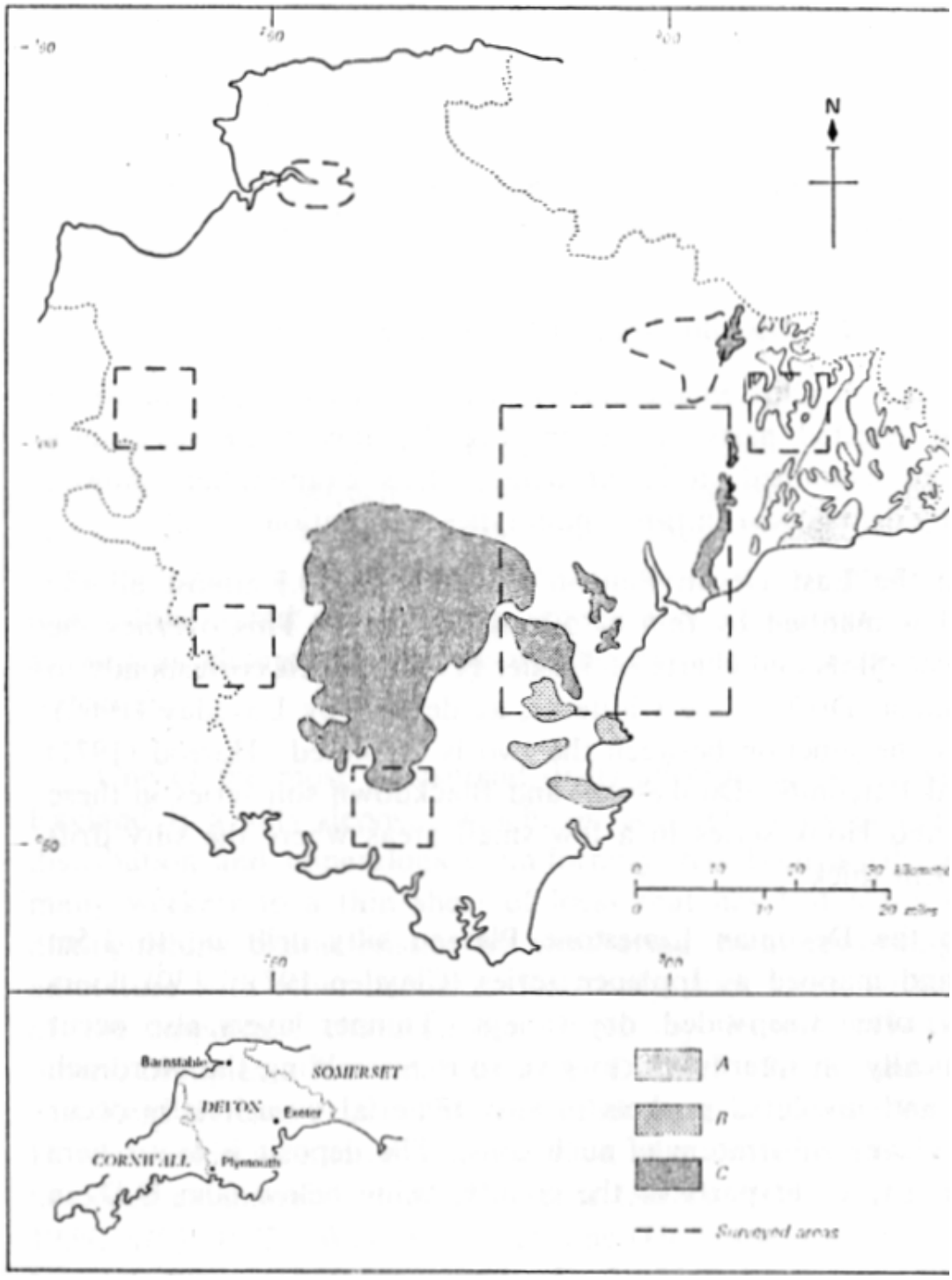

FIGURE 1. Distribution of silty drift deposits in Devon.

A. East Devon Plateau : continuous mantle

B. Devonian Limestone Plateau : valley floors and depressions

C. Other areas : limited random occurrences.

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In well-drained soils, such as the Ipplepen and Nordrach series, the silty drift beneath organically-darkened surface horizons is uniformly yellowish red (Munsell colour 5YR 5/6) to strong brown (7.5YR 5/6). In moderately well drained soils (Batcombe and Hook series) the silty horizons are mostly yellowish brown (IOYR 5/4 to 5/6), and in poorly (Dunkeswell) or very poorly (Blackdown) drained soils they are light yellowish brown (10YR 6/4) or light brownish grey (10YR 6/2) with rusty ferruginous and black manganiferous concretions. All the silty drifts are non-calcareous and slightly stony. The stones consist of flint and chert on the East Devon Plateau, quartz and quartzite pebbles in the drift over Budleigh Salterion Pebble Beds, shale fragments in the Ipplepen soils, and granite fragments on Dartmoor. All these are derived from rocks immediately below, so that the silty drifts must have been partly mixed with the materials on which they rest. However, the stones are often smaller and more angular than those in the deposits beneath, especially on the East Devon Plateau, where the flint and chert fragments lack the continuous patina of similar stones in the Plateau Drift.

Elsewhere in Devon there is no evidence of silty drifts. The main land- units that are devoid of such deposits are the broad outcrops of Devonian and Carboniferous slates and shales north and south of Dartmoor and the Permo-Triassic vales in east Devon. Detailed mapping near Honiton and Newton Abbot showed that the margin of the silty drift coincides closely with the edges of the East Devon Plateau and Devonian Limestone Plateau respectively.

3. Laboratory studies

Eight samples of silty drift from sites over various substrata were selected for analysis. The prefixes A, B and C refer to the areas in Fig. 1.

A1 Dunkeswell (Grid Ref. ST 132065), 30-45cm below the surface, over Plateau Drift (as defined by Loveday, 1962) (Batcombe series).

A2 Dunkeswell (ST 126062), 28-40cm, over Plateau Drift (Blackdown series).

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A3 Dunkeswell (ST 127064), 80-90cm, over Plateau Drift (Hook series).

B1 Ipplepen (SX 838674), 60-95cm, over Devonian Limestone (Ipplepen series).

B2 Ipplepen (SX 835678), 18-46cm, over Devonian Limestone (Nordrach series).

C1 Lukesland (SX 645576), 30-45cm, over granite head.

C2 Little Haldon (SX 919767), 13-58cm, over Haldon Gravels (variant of Southampton series).

C3 Lympstone Common (SY 040854), 120-160cm, over Budleigh Salterton Pebble Beds.

Air dried 10g sub-samples were treated with hydrogen peroxide to remove organic matter, then shaken overnight in dilute sodium hexametaphosphate solution to aid dispersion, and their particle size distributions were determined by sieving and pipette sampling. The detailed size distribution between 4 and 62µm in samples A3, BI, CI and C3 was determined on 2.5g sub-samples, similarly dispersed, using the apparatus described by Stairmand (1950). Fine sand (60-250µm) and coarse silt (20-60µm) fractions were separated from 25g sub-samples by sieving and repeated settling in water respectively. They were then divided into light and heavy fractions with bromoform (S.G. 2.9) and analysed mineralogically with a petrological microscope.

Although the samples are composed mainly of silt (2-62µm), they also contain 15-25% clay and 7-22% sand (Table 1). The amounts of clay are similar to those in partly weathered loess at Pegwell Bay, Kent (Weir *et al.* 1971), but the sand contents are greater, especially in samples C2 and C3. At these two localities the drift overlies sand substrata, so that the incorporation of material from beneath may explain the increased sand content. The main peak of the particle size distribution occurs in the coarser part of the silt range in all samples, but seems to be slightly nearer fine silt in samples from sites west of the Exe valley than in those from east Devon. This was confirmed by the detailed size analyses of silt fractions ; in the samples from East Devon (A3 and C3) the modal value of the coarse silt was 36µm, but in the samples from further west (BI and CI) the value was 26µm).

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TABLE 1. Particle size distribution of loess from Devon

		A1	A2	A3	B1	B2	C1	C2	C3
		ST132065	ST126062	ST127064	SX838674	SX83567	SX645576	SX919767	SY040854
		Batcombe series	Blackdown series	Hook series	Ipplepen series	Nordrach series	over granite head	Variant of South-ampton series	over Bud-leigh Salterton Pebble Beds
		30-45 cm	28-40 cm	80-90 cm	60-95 cm	18-46 cm	30-45 cm	13-58 cm	120-160 cm
CLAY	<2µm	18.4	19.6	20.1	24.8	20.8	15.1	19.6	22.2
	2-41µm	5.0	5.0	4.3	4.9	10.8	7.1	4.2	4.4
	4-8µm	7.2	6.9	5.7	8.5	11.2	10.2	6.5	6.4
SILT	8-16µm	14.0	13.1	10.4	11.4	14.3	15.8	11.5	10.9
	16-31µm	24.4	22.8	22.5	22.6	19.8	22.3	19.1	18.2
	31-62µm	23.2	22.1	23.2	20.6	15.9	15.7	16.8	18.7
	62-125µm	2.3	3.5	5.2	2.1	2.5	2.3	3.0	7.9
	125-250µm	1.7	2.1	3.1	0.9	1.5	2.6	2.7	6.1
SAND	250-500µm	1.4	1.5	2.2	1.1	1.1	2.0	3.9	4.3
	500-1000µm	1.6	2.2	2.2	1.4	1.5	2.5	7.4	0.7
	1000-2000µm	0.8	1.2	1.1	1.7	0.6	4.4	5.3	0.2

The coarse silt fractions of all samples are mineralogically alike. Their light fractions contain 79-85% quartz, 10-16% felspar, and uniformly small amounts of muscovite, glauconite and flint (Table 2). The amounts of opaque heavy minerals (mainly iron oxides), ranging from 0.1 to 2.0% of the coarse silt, are probably dependent on soil drainage conditions, but the quantities of non-opaque heavy minerals are less variable (0.8-1.2%). The composition of the non-opaque heavy fraction is also very constant (Table 2), and emphasises the uniformity of the silt in the samples.

In contrast, the fine sand fractions vary more in mineralogical composition. In samples Al, A2 and A3 they are essentially similar to the coarse silt fractions, except that felspar is less common and flint a little more common. BI and B2 contain the same sand mineral assemblage, but with the addition of shale fragments and chalcedony, together constituting 20-25% of the fine sand. The chalcedony is probably derived from "beekite" in the underlying Devonian Limestone (Wickes 1910), and the shale fragments from shale bands interbedded with the limestone. In sample C1 the fine sand contains 40% alkali felspar, >2% tourmaline and >1% biotite, indicating the incorporation of material from the granite. Similarly, the fine sand of C2 contains much flint, probably derived from the underlying Haldon Gravels, and a heavy mineral assemblage with topaz and much more tourmaline and biotite than occurs in the coarse silt. Finally, comparison of the fine sand mineralogy of C3 with that of a Budleigh Salterton Pebble Beds sample from Harpford Common (SY 055900) and with Thomas's account (1902) of detrital minerals in the Pebble Beds, showed that no more than 20% of the sand could have come from the Pebble Beds.

Therefore, on the basis of their fine sand mineralogy, the samples are divisible into two groups. Those from the west (B1, B2, CI, C2) contain fine sand that is mineralogically different from the coarse silt, whereas those from east of the Exe (Al, A2, A3, C3) contain fine sand that is mineralogically much more like the coarse silt. In the second group the silt is coarser, and much of the fine sand is in fact the coarse tail of this component.

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TABLE 2. Mineralogical composition of 20-60μm fractions from Devon loess samples

	A1	A2	A3	B1	B2	C1	C2	C3
	ST132065	ST126062	ST127064	SX838674	SX83567	SX645576	SX919767	SY040854
	Batcombe series	Blackdown series	Hook series	Ipplepen series	Nordrach series	over granite head	Variant of South-ampton series	over Bud-leigh Salterton Pebble Beds
	30-45 cm	28-40 cm	80-90 cm	60-95 cm	18-46 cm	30-45 cm	13-58 cm	120-160 cm
A. LIGHT FRACTION								
Quartz %	82	82	82	85	83	79	81	81
Alkali Felspar %	14	14	14	10	14	16	14	14
Muscovite %	2	2	2	1	1	3	2	3
Glauconite %	1	1	1	2	1	1	1	1
Flint and chalcedony %	1	1	1	2	1	1	2	1
B. HEAVY FRACTION (NON-OPAQUE)								
Chlorite ‰	595	566	565	574	397	522	538	640
Biotite ‰	20	23	29	40	23	45	23	50
Epidote ‰	137	150	162	170	189	156	178	98
Zoisite ‰	15	8	5	4	24	12	12	9
Zircon ‰	75	75	88	76	104	36	57	76
Tourmaline ‰	36	41	38	42	53	80	51	38
Yellow Rutile‰	25	23	33	20	38	11	14	19
Brown Rutile ‰	7	6	8	7	11	4	6	6
Red Rutile‰	2	-	3	-	2	2	2	1
Anatase ‰	19	24	19	17	26	9	12	26
Brookite ‰	3	2	5	-	6	3	1	3
Green Hornblende‰	35	37	22	27	63	68	59	16
Brown Hornblende‰	2	4	4	2	4	5	5	3
Tremolite ‰	11	14	6	7	19	13	17	10
Garnet ‰	12	20	5	11	32	25	16	1
Kyanite ‰	1	2	3	2	3	1	3	3
Staurolite ‰	5	5	5	1	6	8	6	1
Number of grains counted	1105	837	1092	1052	1229	1105	1035	685

4. Discussion

Mineralogically uniform silt could only have been widely distributed as a thin veneer over the higher parts of south and east Devon by wind. It is therefore probable that the silt and mineralogically similar fine sand were derived from distant sources. The locally derived coarser parts of the drifts (sand and stones) could have been mixed with the aeolian silt by various processes, such as surface movement of material by rainwash during deposition of the silt, or frost working and biological mixing of the soil after deposition.

The aeolian component of the drifts is slightly coarser in east Devon than further west. This suggests that the silt was blown from the east, with its coarser, heavier particles settling slightly sooner than the finer, lighter ones. If allowance is made for the incorporation of the coarse, locally derived material, the particle size distribution of the silty drifts is similar to that of loess, so that comparison with the loess deposits of eastern England may help to establish the age and origin of the Devon drifts. The coarse silt fractions of the loess at Pegwell Bay (Kent) and the Norfolk coverloams are both composed of the same minerals as the silt in the Devon samples, but contain more felspar, flint, epidote, zircon, garnet and amphibole, and less quartz, muscovite, chlorite, biotite and tourmaline (Catt *et al. 1971,* Table 3). These differences may indicate separate sources for the loess in the two areas. However, if the Devon loess was derived from the east, the differences could also result partly from the winnowing effect of the wind, the platy, micaceous particles being carried further than the more equidimensional particles. The composition of the Chilterns loess analysed by Avery *et al.* (1969 Table 5.4 ; 1972 Table 6) supports this suggestion, because it is intermediate in both geographic position and mineralogical composition between samples from the eastern counties and those from Devon, but more analytical data on samples from intervening areas is needed before this lateral correlation is firmly established. If correct, the correlation would indicate that the aeolian silt in Devon was deposited in the later part of the Last Glaciation, probably during the Upper Pleniglacial of van der Hammen *et al.* (1967), and was derived at least in part from glacial outwash deposits in the

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North Sea basin (Catt *et al.* 1971). The loess may be similar in age to the upper Head of the south Devon coast, as this often contains silty material attributed to loess (Mottershead 1971).

The shattered stones in the silty drifts and the involutions at the junction of the silty drift with Plateau Drift or the clayey substratum on the Devonian Limestone Plateau both suggest periglacial frost working during or after the later stages of loess deposition. Periglacial disturbance, as expressed by involutions, was rare in south-west England during the Last Glaciation, especially at levels as low as the Devonian Limestone Plateau (Williams 1969). Consequently it would have been most likely to occur during the coldest phase, which probably coincided with part of the Upper Pleniglacial.

ACKNOWLEDGEMENTS. We thank Mrs. E. A. Merriman for drawing the diagram. B. Clayden, D. V. Hogan, D. J. R. Manley and S. J. Staines did much of the fieldwork on which the diagram is based.

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THE FLANDRIAN SEA-LEVEL RISE IN THE BRISTOL CHANNEL

by C. Kidson and A. Heyworth

Abstract. Submerged and buried forest peats, from the Bridgwater Bay area of Somerset are described. Deductions are made. from pollen analyses, regarding their heights of formation, relative to contemporaneous sea-level. Sedimentation is seen to have kept pace, throughout the Flandrian, with the rise in sea-level, and to have occurred up to a height midway between M.H.W.O.T. and M.H.W.S.T. Most peats have been formed very closely around this height, which is also that at which the prehistoric wooden trackways of the area were constructed.

Radiocarbon dates from these peats are used, after applying corrections for subsequent gravitational compaction, to produce a eustatic sea-level curve, for the last 10,000 years. The area appears to have been isostatically stable throughout this period. A second sea-level curve is produced, after dendrochronological correction of the radiocarbon dates, which is exponential over the whole 10,000 years. No sea-levels higher than the present are indicated. Such levels elsewhere, and the oscillations in other published sea-level curves, are attributed to isostatic effects and periods of varying storminess.

The "Romano-British Transgression" is discussed, but no evidence can be found to justify regarding it as a discrete phenomenon.

1. Introduction

The intertidal zone in Bridgwater Bay contains well-developed submerged forest beds, which are exposed over an unusually extensive area. This is due to the large tidal range (12m at springs), and to the nature of the intertidal deposits. Pollen analysis and radio-carbon dating of the peats of the submerged forest beds and of corresponding peats from a large number of boreholes (Figure 1) have been carried out. Micro-faunas from the intercalating marine estuarine and freshwater clays have also been examined. A principal object has been to relate the height of peat formation to the levels of the transgressing Flandrian Sea. The results have been used to construct a sea-level curve covering the last 10,000 years.

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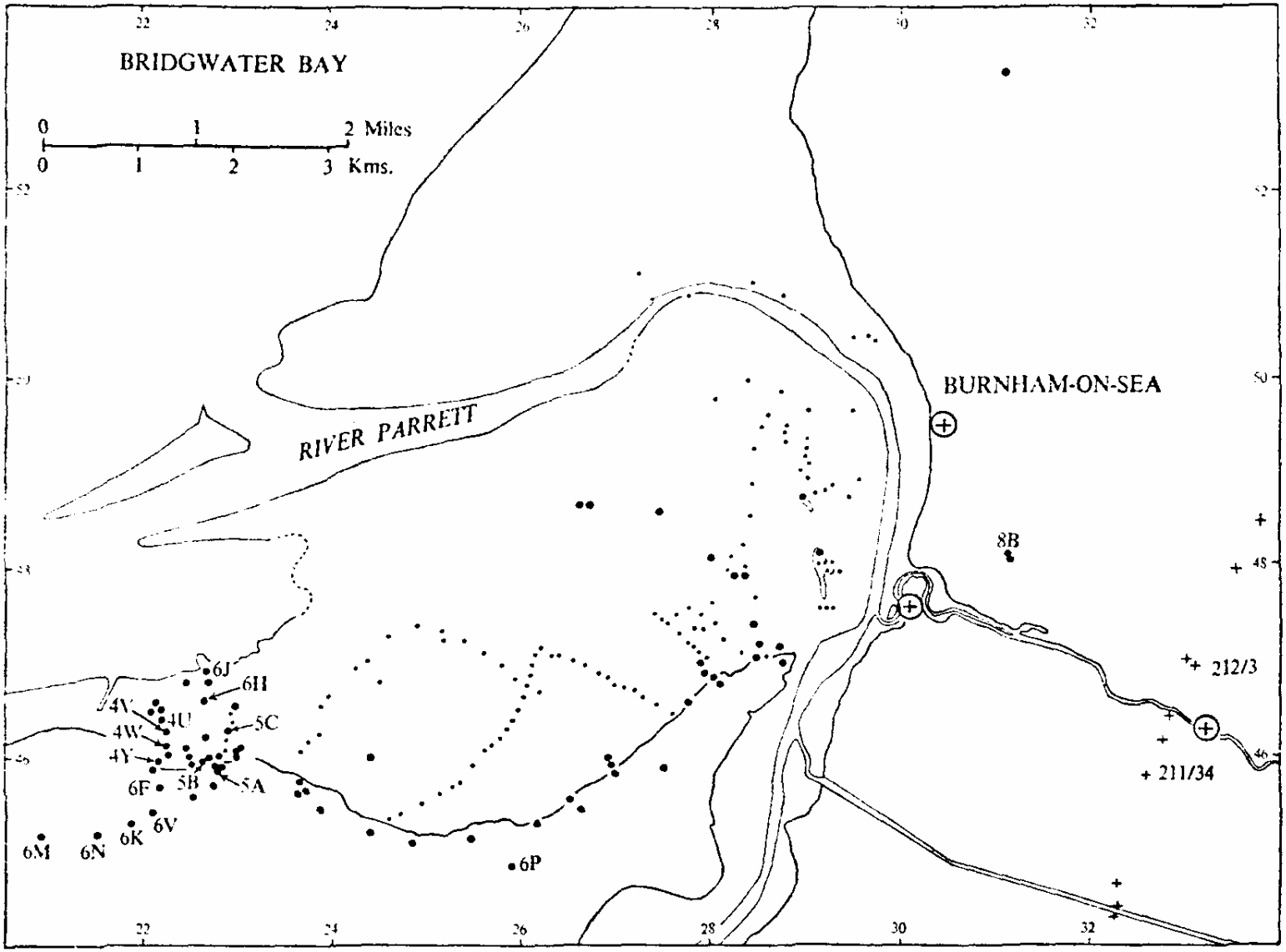

FIGURE 1. Bridgwater Bay. Boreholes and probes.

Boreholes put down by the present authors (1962-69) are indicated by large dots, probes by smaller dots.
Boreholes sunk by Soil Mechanics Ltd. for the M5 Motorway are shown by small crosses.
Boreholes sunk for the Somerset River Authority are shown by larger crosses.

Several such curves have been published in recent years, using data from many areas of the world. They include those of Jelgersma (1961, 1966), Fairbridge (1971), Shepard (1963) and Mörner (1969). These curves fall into two classes. The first (Jelgersma and Shepard) shows a smooth curve rising continuously, but at a decelerating rate, to the present day. The second (Fairbridge and Mörner) suggests an oscillating sea-level, with an overall rise, which reached present sea-level by 3700 B.C. (Fairbridge) to 2000 B.C. (Mörner), and subsequently exceeded it.

Of the above mentioned curves, that of Mörner would appear to be the most reliable in that isostatic effects appear to have been largely eliminated, whilst the other studies are not so corrected. The curve here presented is in general agreement with that of Morner, although it resembles those of Jelgersma and Shepard in showing a smooth and continuous rise to present sea-level. It suggests that the oscillations which Morner envisages may not all be real eustatic effects.

The data used by Fairbridge are from several continents and the oscillations in his sea-level curve can be shown to be largely the result of differing isostatic effects (Jelgersma 1966). Jelgersma's (1966) curve is based on data from the Netherlands, an area which has suffered long-term subsidence, and the curve must, therefore, be somewhat below the true eustatic one. Shepard (1963) used data from world-wide areas supposed to be relatively stable. In avoiding areas of uplift, however, it appears that he has obtained an average of slight subsidence, over the last 7000 to 8000 years, his dates for this period being, again, mainly from the Netherlands. The upper part of Shepard's curve, therefore, whilst above that of Jelgersma, is probably still too low.

Many radio-carbon dates and pollen diagrams have been published from the Bristol Channel and Somerset Levels area, especially by Godwin and his co-workers. They published an eustatic curve, including dates from the Somerset Levels (Godwin, Suggate and Willis, 1958) and interpreted it as indicating that the present sea-level was reached 5500 years ago. The basis of this conclusion was the date from Tealham Moor (Q120) of 5412 ± 130 B.P. for a peat at O.D. It is, however, apparent that sedimentation and peat growth in such localities is related, not to mean sea-level,

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as implied by Godwin *et al.* but to a height between that of high-water neap tides and high-water spring tides. When the large tidal range in the Bristol Channel is taken into account, the Tealham Moor date indicates a rise in sea level, neglecting compaction effects, of about 4m. *since* the formation of this peat. A curve drawn from data largely corresponding to that used by the present authors has been published by Hawkins (1971). Although the Bristol Channel is considered to be a stable area, this curve is well below that presented by Jelgersma (1966) for the Netherlands. Hawkins used six dates obtained by the present authors from samples taken from their boreholes (Figure I) without consulting them on their use or interpretation. Their own interpretation differs significantly from his.

The question of crustal stability in the Bristol Channel area has been discussed by Godwin (1943 and 1956) and Churchill (1965). Both these authors considered that it has been a relatively stable area. Churchill sought to demonstrate, using samples taken from peats laid down 6500 years ago, from sites between the Irish Sea and the Netherlands, that a tilt from north-west to south-east has taken place since that time. He assumed, from world-wide data, a mean sea-level at 6500 B.P. about 3m lower than at present. The three samples from the Bristol Channel which he considered, at Margam, Westward Ho! and Burnham-on-Sea, are all still very close to the height appropriate to this sea-level. When subsequent compaction and differences in heights of formation relative to sea-level are considered, the discrepancies become negligible. Any tilt would therefore seem to hinge in the Bristol Channel area which appears to have been unaffected by any isostatic or tectonic movement in the Flandrian. Despite the theoretical calculations of isostatic movement recently published (Walcott 1972) the area is regarded as essentially stable.

An attempt has been made to eliminate errors in sample height due to gravitational compaction and the present curve, is, therefore, taken as representing the eustatic sea-level rise. All the dates used are from the same small area, and the assumptions regarding its stability appear to be borne out by the results.

2. Sampling

The exposed submerged forest, whilst often difficult of access is easily sampled. In other areas it is covered by mud or clay,

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or both, and hand-augering had to be resorted to. Over the great majority of the landward area the peats are covered by estuarine and fresh-water clays, and power-augering has been required for most of the boreholes. The samples from the authors' own boreholes were supplemented by others obtained from the bore-holes put down on the line of the MS Motorway, (Figure 1). Almost all the peat samples have been subjected to pollen analysis, and, in addition, many of the clays have been examined for foraminifera.

3. **Types of Peat**

The pollen and macro-fossils show that the peats were formed from vegetation ranging from salt-marsh to forest. All types of peat are found at all heights, and there appears, at first, to be no pattern. Peats and clays outcrop on the beach, in a manner suggesting repeated transgressions and regressions. Pollen and C14 dating showed that peats were successively younger, passing up the beach. However, recent stripping off, by storms, of large areas of mud, has shown that many of what appeared to be separate peat layers are, in fact, outcrops of a single extensive sheet. This covers the undulating beach slope from shortly below the storm beach down to extreme low water mark, and beyond. It is absent in some places, apparently as a result of removal by wave-attack. Large rafts of peat cut away from the sheet, in this way, can be seen on the beach at the present time. The age of the peat ranges from c. 3000 B.P. just in front of the storm beach to c. 7000 B.P. at L.W.M.S.T.

The sheet rests, very largely, on solid or weathered Lias, and it can be correlated with the basal peat found in boreholes to landward. This is found down to - 20m O.D., and forms a more or less continuous cover over the buried Pleistocene surface (Figure 2). Samples from any area of this peat, from the same height are found, consistently, to be of the same age. This suggests that the height of formation of this peat was related to sea-level. The picture is of a rising water-table, resulting from the rise of sea-level, creeping up around the previously well-drained woodland, and creating a margin of swamp and carr. In this belt, peat formation would be initiated, and continue until overtaken by increasing salinity. Thus, a blanket of basal submerged-forest peat would be developed over large areas of undulating land surface.

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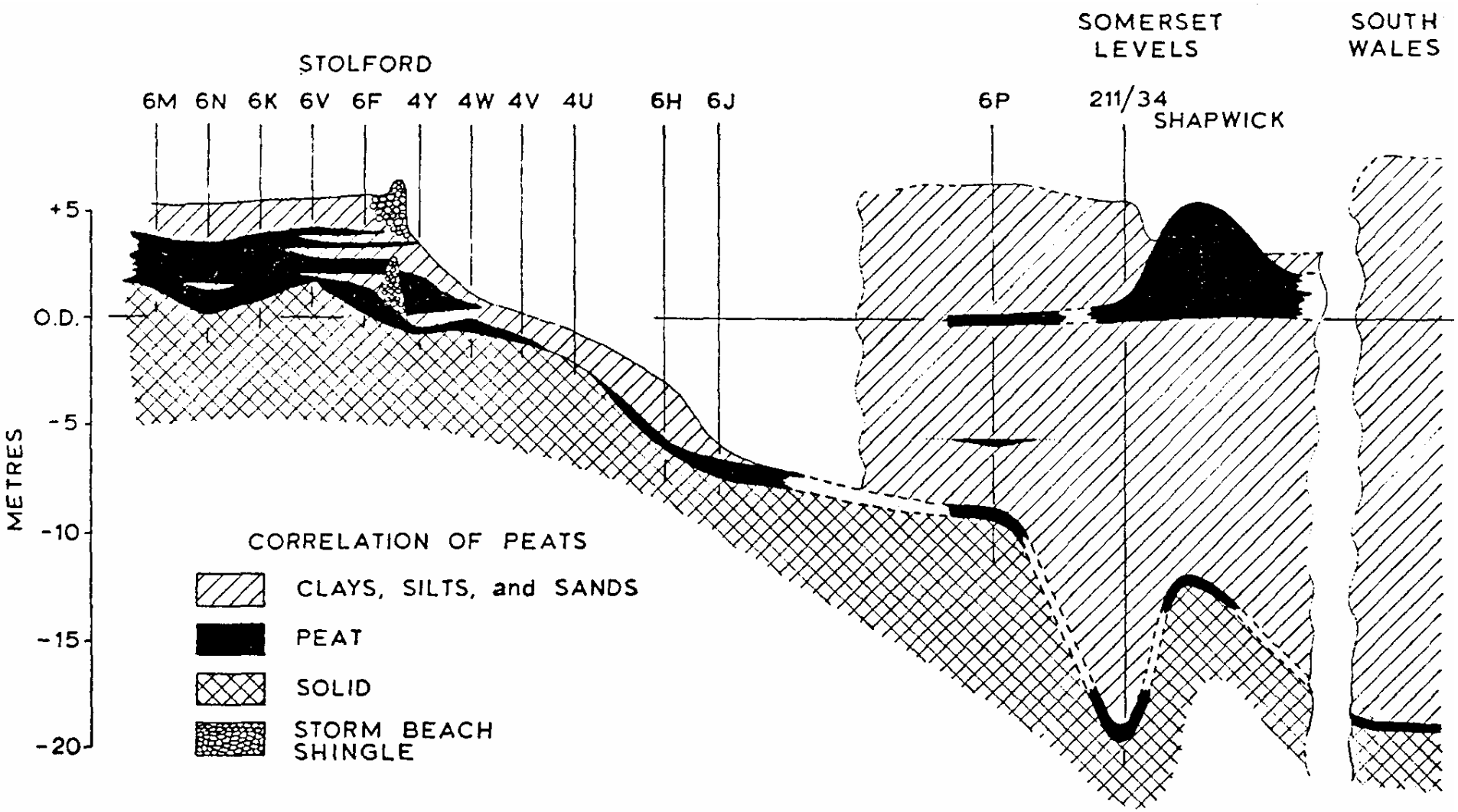

FIGURE 2. Diagrammatic section of the Flandrian deposits of the Bridgwater Bay and Somerset Levels area. The correlation suggested with South Wales relates to Margam and Baglan Burrows (Table I). The break in the section is made to draw attention to the deep buried channels of the Severn and its tributaries (Anderson 1968) which intervene in the Bristol Channel.

In most cases it appears that, probably as a result of protection behind a storm beach, the peat so formed was covered by clay before being reached by the zone of wave attack, and was thus preserved.

From the foregoing it can be seen that the age of the basal peat in any section serves to date contemporaneous sea-level. The height of the peat appears to represent a height very close to "M.H.W.O.T." It is highly unlikely, however, that the process was a smooth and continuous one. A large part of the marine transgression was doubtless accomplished during periods of storminess and by storm surges. In the intervals between these atypical episodes there was a return to more normal conditions. This pattern would explain the many minor short-term oscillations seen in the sea-level curves of some investigators.

At any one site, after the formation and inundation of the basal peat there is seen to have been a period of intertidal or estuarine clay deposition, during which the formation of the basal peat was going on at higher levels elsewhere. These events continued until the rate of sea-level rise had slowed to such an extent that it began to be matched by the rate of sedimentation, allowing recolonisation of coastal clays by peat-forming vegetation. During this period, any peats formed are found to be thin, mainly salt-marsh types, resulting from minor local prograding episodes.

At about mean sea-level (O.D.) a marked change is seen in almost all sections, shown by the termination of blue estuarine clay deposition. On the very extensive clay thus formed a second submerged forest peat is often developed. In the inland parts of the Somerset Levels peat growth has been continuous from this time onwards, giving rise to extensive raised bogs. These have obviously grown to heights well above the surrounding water-table and cannot be related to past sea-levels. No dates from ombrogenous peats have been used, therefore, in constructing the sea-level curve.

In the more coastal areas, the peats above the "O.D. clay" surface are intercalated with fresh water or brackish clays, and are, over almost all the area, covered by a topmost layer of floodplain clay, which forms the present land surface. These peats can be seen to have been formed very close to contemporaneous water-table and can, therefore, be used to date sea-level.

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4. Heights of Peat Formation relative to Sea-Level

The disadvantages of coastal peats as material for pollen-analysis are well known, and the Bridgwater Bay peats show them no less than others. The peats often have a very high clay or silt content, which is difficult to remove. In the intervening clays, even when these are highly organic, pollen is so scarce and the pollen spectrum so distorted that attempts at pollen counting are largely fruitless. In boreholes it is common to encounter large trunks or stumps, resulting in sections completely devoid of pollen.

Local influences are so preponderant in determining the nature of the peats that entirely different types may be found in very small areas, and correlation is extremely difficult. Climatic effects, and the general forest composition, are almost completely concealed, particularly as the peats are likely to be encountered as thin, isolated sections. However, such local effects are advantageous in distinguishing the rapidly changing environments and in making possible deductions about the heights of peat formation relative to sea-level.

Pollen diagrams showing these effects are included here, from the two main types of sequence encountered in the coastal deposits. 6N is a borehole through thick peats in the centre of a small basin in the Lias. This basin, though open to the sea, has a rock floor at about -2m O.D., and is infilled to about O.D. by freshwater sands and clays. Salt water could not therefore, have had access to the basin during the earlier, rapid period of eustatic rise. Borehole 6P, on the other hand, passes through largely estuarine clays, where deposition has taken place more or less continuously since the sea reached the level of the weathered Lias surface at about - 10.5m O.D. Here, the periods of freedom from salt water have been short, and the peats are thin, and widely separated.

In the 6N Tree-Pollen diagram (Figure 3) the variations in the high *Alnus* levels largely obscure the regional pattern. However, the preponderance of *Alnus* pollen is an indication of the local dominance of the species. Whole anthers and catkins of *Alnus* occur in the peats, whilst its wood is also very common. This maintained presence indicates a long-standing alder carr environment. This cannot be explained by climatic influences, and

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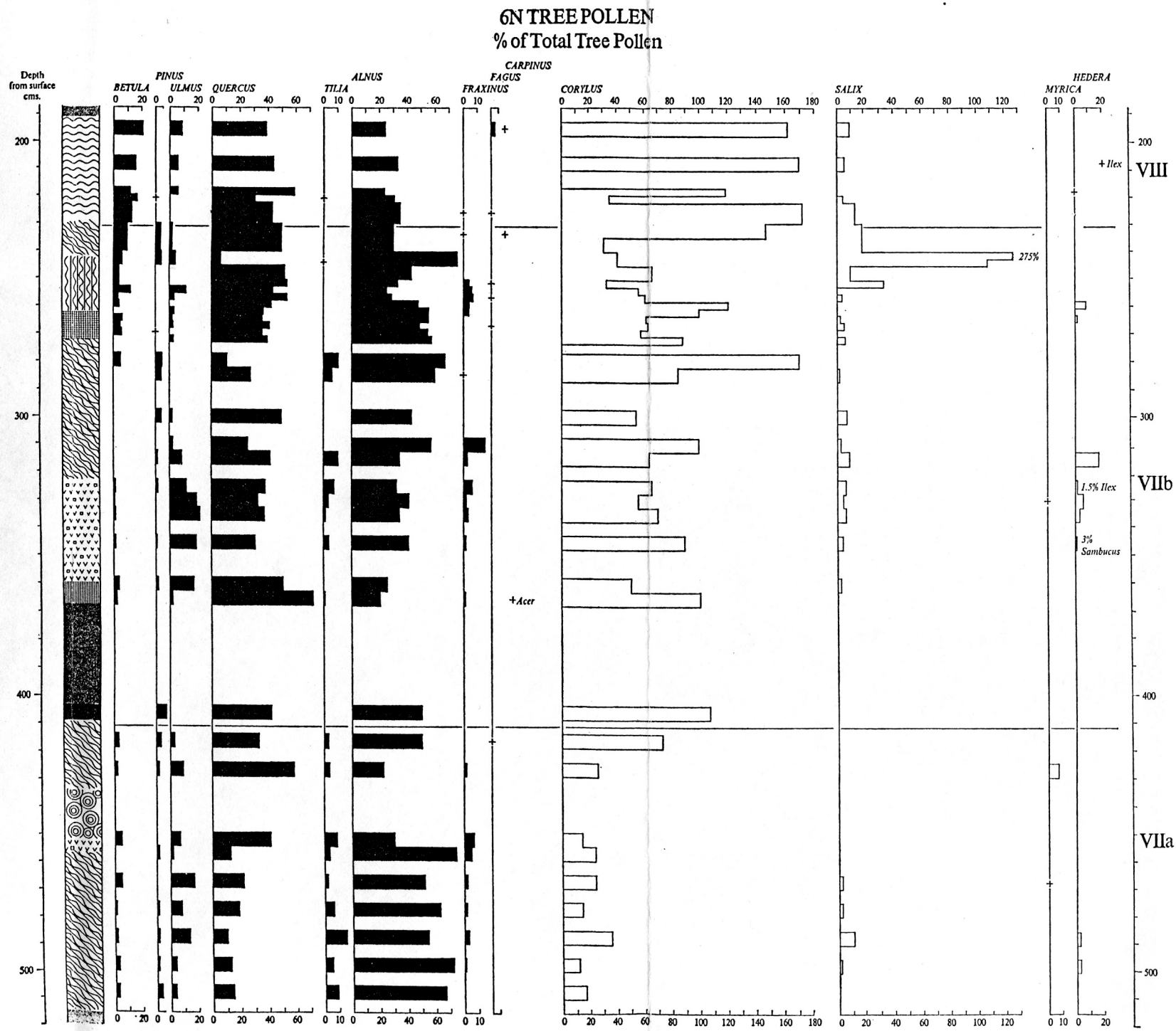

FIGURE 3. Borehole 6N (Figure 1) Tree Pollen.
(for explanation see text).

must be the result of a continuously high water-table, constant relative to the growing peat surface, through a vertical distance of almost 3m. As the basin was, throughout this period, open, and draining to the sea, it is clear that this rise of water-table must be related to the rise in sea-level. It is noteworthy that similar peats in the submerged forest exposures to seaward show maintained alder pollen values of over 90% of tree pollen, and 75% of total pollen. This suggests that at the 6N site, drainage water contributed to a more eutrophic environment than obtained outside the basin.

The high *Salix* peak is also an indication of the general wetness, marking the end of an open-water phase reflected in the non-tree pollen diagram (Figure 4). Very high levels of desmids and diatoms are succeeded in turn by peaks in *Sparganium, Salix,* ferns and *Corylus.* Other indications of a high water-table throughout are given by the figures for *Phragmites, Cyperaceae, Lastrea* and *Typha.* Chenopodiaceae peaks show a sensitive equilibrium with salt water.

Only above 230 cm, where a sharp change occurs, is there any indication of ombrogenous peat growth above the general watertable, with the sudden appearance of unhumified *Sphagnum* peat. This change, whilst not representing a recurrence surface may be of the same date as the "Grenzhorizont." Apart from this, there is a remarkable absence of any raised-bog species, such as Ericaceae. The open-water phase below the *Sphagnum* peat may have been caused by the blocking of the basin mouth by the storm beach. This, coinciding with the virtual cessation of sea-level rise, would bring about a constant water-table allowing an increase in acidity. Peat growth could then continue only by the development of a raised-bog. Increased rainfall at this date may also have been responsible for the sharp change seen in the peats. The stabilisation of the coastline occurring at this time and the development of coastal vegetation similar to that of today is shown by the decline in tree-pollen and the rise of *Phragmites* to high values. It is clear that, at least below 230 cm., peat growth continuously kept pace with the rise in water-table, and that this rise could have been due only to the rising sea-level.

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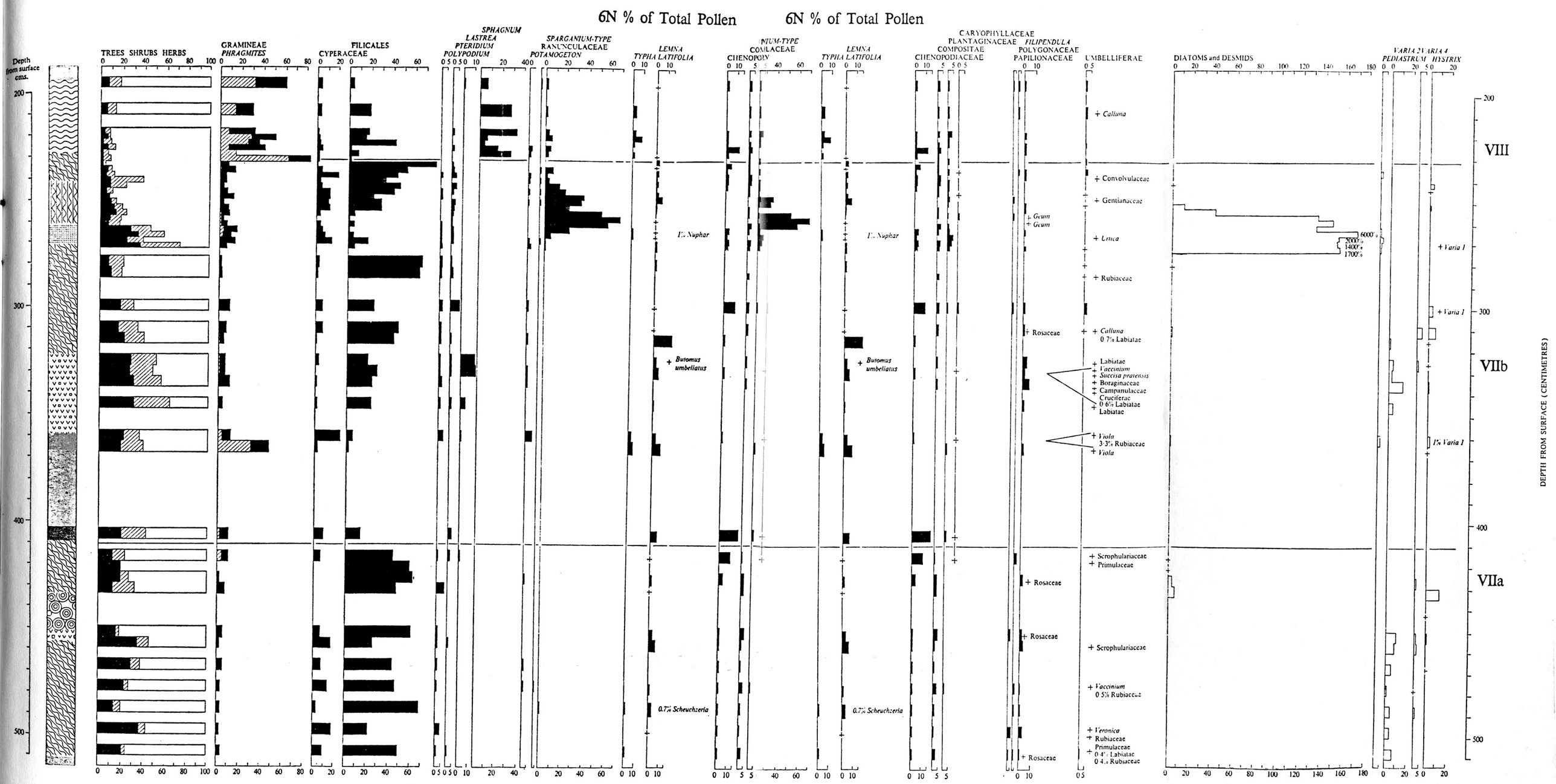

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ehole 6N (Figure 1) Non-Tree Pollen.

orehole 6N (Figure 1) Non-Tree Pollen.

In the non-tree pollen diagram for borehole 6P (Figure 5) the influence of the consistently high water-table is again seen in the levels of *Cypernceae, Lastrea, Sparganium, Typha* and *Lemna.* Brackish and salt-marsh conditions are indicated by the high Chenopodiaceae values, the presence of *Althaen* and 'microforaminifera' and the diatom species. There is much less tree-pollen of local origin than in borehole 6N, *Alnus* being dominant only from 1190 to 1205 cm., and being almost absent from 1500 to 1545 cm. The site appears to have been, throughout, at, or beyond, the seaward margin of an alder carr.

There are obvious variations, in this, as in other boreholes, in the conditions of peat formation. This is illustrated by the two radiocarbon dated samples. 1-2689 (6890 ± 120 B.P.) is from a peat formed largely of the grasses and sedges of a high marsh, stabilised and built up above the surrounding level, allowing the initiation of peat formation. Fresh water is later seen to have been ponded up, as shown by the immediately subsequent peaks in *Cyperaceae, Lastrea* and *Sparganium* values. The thinness of the peat shows, assuming no fall in sea-level, that the height of formation of the samples above the water-table could not be more than about 20 cm.

I-2690 (7360 ± 140 B.P.) is from a peat containing very little pollen derived from the immediate locality, *Ulmus* and *Quercus* being preponderant. It is clear that it accumulated at the bottom of shallow, open water, the bulk of the peat consisting of laminated leaf remains. Some fen species are present and suggest that the depth of water was probably about 20 cm.

Pollen analysis of all the other coastal peats encountered in Bridgwater Bay produces results similar to the above, raised bogs being absent, and the influence of a high water-table being the dominant factor. In a few cases, e.g. Borehole 5A., C14 sample N.P.L. 146, plant remains are found in low salt marsh peaty clays, formed at heights estimated as up to 1.2m below " M.H.W.O.T."

It is assumed throughout, that the major factor in determining the upper limits of sedimentation and peat formation was the drainage base-level. This, in turn, is assumed to have been determined by sea-level, the "effective water-table" being maintained at about "M.H.W.O.T." It is not, of course, envisaged

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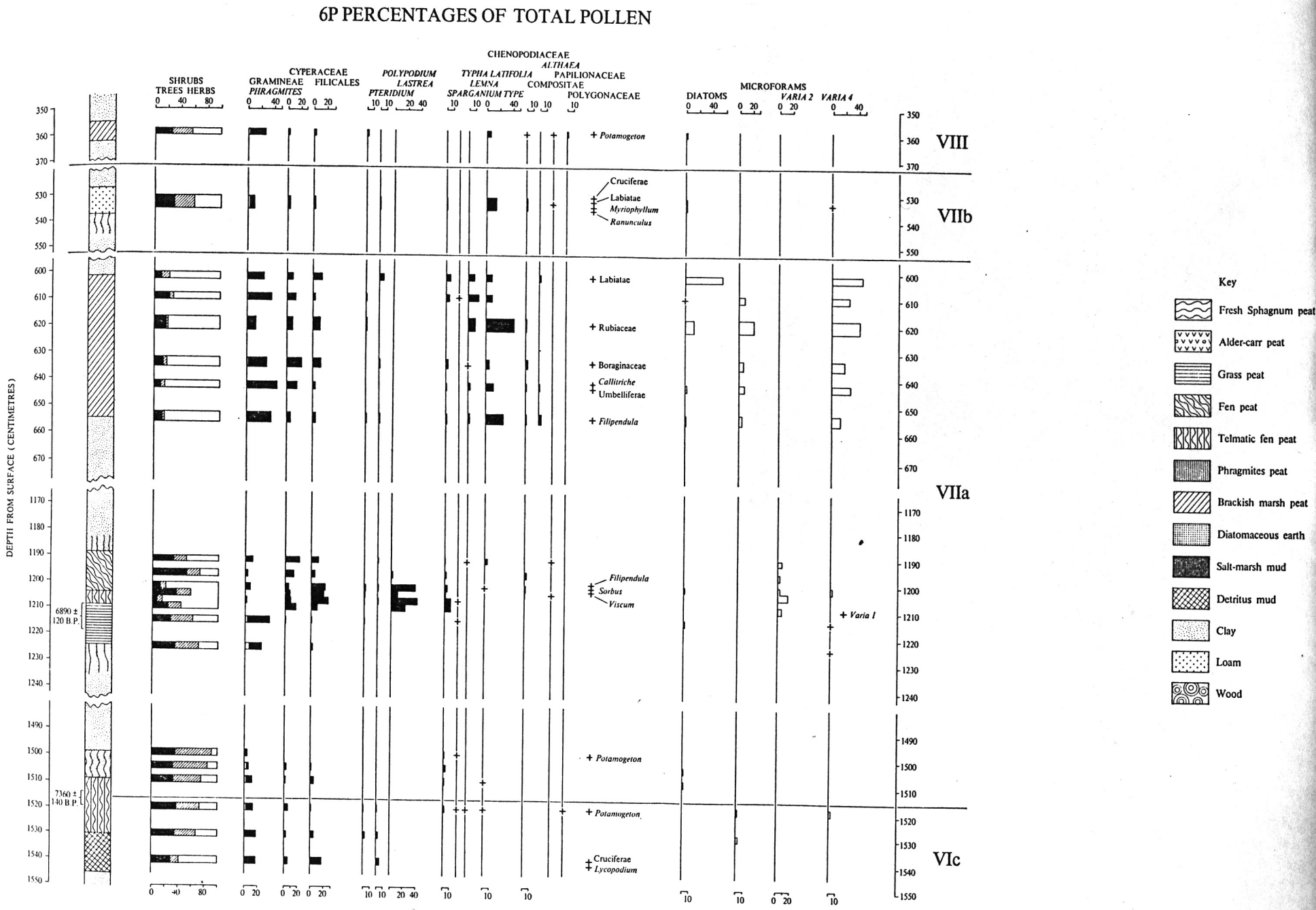

FIGURE 5. Borehole 6P (Figure 1) Non-Tree Pollen
(for explanation see text).

that the actual water-table would remain exactly constant at different seasons and stages in the tidal cycle. A similar zonation of vegetation as is seen at the present day would, however, exist, and bear the same relationship to the `effective water-table'; moving upwards with the rise in sea-level. Appropriate peat-forming communities would become established in accordance with the differing periods of salt or fresh water submergence. Only in the case of trees would the rate of sea-level rise outstrip the rate at which the communities would change, and so cause deviations from this equilibrium. The above assumptions are consistent with the fact that the peats other than the basal sub-merged forest rest on essentially horizontal clay surfaces. Exceptions to this are seen only where the Pleistocene topography rises above the "O.D. clay" surface, when the inclined submerged forest continues up the slope and merges with the higher peats.

A further assumption is that the tidal range in the estuary has remained constant throughout the period considered. There can have been very little change during the last 6000 years. Between 10,000 and 6,000 years ago, there may have been slight changes, as a result of the entry of the rising sea into a widening channel. These may not have been as marked as might at first appear, however, as sedimentation and infilling have largely kept pace with the rise in sea-level.

5. Prehistoric Wooden Trackways

Dates obtained from these trackways (Godwin 1960) have been used in plotting the sea-level curve. The construction of these trackways has not previously been considered to have resulted from the effects of a rising sea-level, but rather to changes in the water-table brought about by rainfall variations. However, it is difficult to see how the water-table in the areas concerned could, whatever the rainfall, have been substantially different from that elsewhere in the vicinity. The drainage base-level must have been determined by sea-level, as the peats concerned all rest on estuarine clay and no isolated basins are involved. Godwin has, in fact, suggested that the uppermost flooding horizon in some peats (not associated with trackway construction) might be the result of ponding-up of freshwater as a result of the "Romano-British Transgression" (Godwin 1948: 279). He gives the estimated

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date of this horizon as A.D. 150, and this is plotted on the sealevel curve (Figure 7). The deterioration in climate, and increased rainfall, with a slowing of peat growth, after the Climatic Optimum may well have been factors determining the precise dates at which peats were overtaken by the water-table.

No corrections for subsequent compaction, or for differences in height relative to the water-table have been made in respect of the C14 dates from the trackways. The piles were driven into the peat, and the rest of the timber and brushwood used was presumably well-consolidated at the time of construction and by subsequent traffic. Any gravitational compaction since then must be negligible. Similar reasoning suggests that the trackway timbers must have been very close to the water-table. Many complications could be introduced here, since the dates are not those of trackway growth, but of the trees used in the construction. However, the conclusions of the trackway excavators were that the great majority of the timber was felled nearby and used immediately. The height at which the trees grew is, of course, irrelevant in this instance. Dates from the centres of large trunks would, as elsewhere, be very misleading.

6. South Wales Dates

South Wales is sufficiently near to make any great differences unlikely and dates from sites there have been included in the curve. These are from the Port Talbot area, the first series being from a borehole at Margam (Godwin and Willis 1961). Three dates here (Q265, Q274 and 0275) appear to be from peats formed at heights determined by sea-level. However, the top two (Q265 and Q274) from what must originally have been a very thick peat, have obviously suffered considerable depression, as a result of gravitational compaction. The peat in question is now covered by a very much greater thickness (14m) of sediments than other equivalent peats, and there is reason to think that the compaction corrections applied are, therefore, somewhat doubtful. Q275, from the base of this peat, where the likely error is small, is, as a result, the only one of the three used in constructing the curve (Q275, 6184 ± 143 B.P.).

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TABLE 1. COMPACTION CORRECTIONS at some BRISTOL CHANNEL SITES

Site/Borehole	No. on Curve	Lab. No.	Age in C14 yrs. B. P	Present Ht. (metres O.D.)	Ht. Before compaction (metres O.D.)	Ht. rel. to contemp. water-table (metres)	Ht. of contemp. water-table (metres O.D.)
Viper's Platform	1		2410+110				
		Q311	2460+110	+3.8	Unchanged	0.0	+3.8
Shapwick Heath Trackway	2	Q39	2470+110	+3.8	Unchanged	0.0	+3.8
Westhay Track	3	Q308	2800+110	+3.6	Unchanged	0.0	+3.6
Meare Heath Track	4	Q52	2840+110	+3.8	Unchanged	0.0	+3.8
5A		NPL146	3460+90	+0.6	+1.4	-1.2	+2.6
Honeycat Track	5	Q429	4215+130	+2.4	Unchanged	0.0	+2.4
Blakeway Farm Track	6	Q460	4460+130	+2.4	Unchanged	0.0	+2.4
Honeygore Track	7	Q431	4750+130	+2.4	Unchanged	0.0	+2.4
6V		I 3395	4790+120	+2.3	+2.5	+0.5	+2.0
6F		I 3396	5250+140	+0.8	+1.4	+0.5	+0.9
6F		I 3397	5330+120	-0.5	Unchanged	-0.5	0.0
5B		NPL147	5380+95	-1.0	-0.2	-0.5	+0.3
Tealham Moor		Q120	5412+130	0.0	+0.5	0.0	+0.5
Shapwick Heath		Q423	5510+120	+0.3	+0.5	0.0	+0.5
Tealham Moor		Q126	5620+120	0.0	+0.5	0.0	+0.5
Margam.		Q275	6184+143	-3.2	-1.9	0.0	-1.9
5C		NPL148	6230+95	-2.0	-1.8	0.0	-1.8
Burnham-on-Sea		Q134	6262+130	-4.6	-4.0	-1.0	-3.0
6P		I 2689	6890+120	-5.95	-5.0	+0.2	-5.2
6J		I 2688	7060+160	-7.3	-6.5	0.0	-6.5
8B		I 3713	7320+120	-8.45	-8.4	-0.7	-7.7
6P		I 2690	7360+140	-8.95	-8.2	-0.2	-8.0
212/3		I 4403	8360+140	-21.3	-20.0	-0.5	-19.5
211/34		I 4402	8480+140	-19.5	-19.0	-0.5	-18.5
Baglan Burrows		Q 663	8970+160	-18.9	-18.0	0.0	-18.0

TABLE I. Compaction corrections at some Bristol Channel sites.

Sites 1-7 see Godwin 1960. pp. 6-22.

Sites 5A, 6V etc. dates obtained by present authors from boreholes put down by them.

Sites 212/3, 211/34 dates obtained by present authors from boreholes put down by Soil Mechanics Ltd.

Tealham Moor, Burnham-on-Sea } see Godwin, Suggate and Willis 1958.

Shapwick Heath, Margam } see Godwin and Willis 1961.

Baglan Burrows see Godwin and Willis 1964.

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The second series of dates is from Baglan Burrows (Godwin and Willis 1964). The dates range from 11,980 B.P. to 8,970 B.P., within a height range of only 80 cm, and it is clear that peat growth was here initiated in a local basin, much above sea-level. The peat is the equivalent of the basal submerged forest peat, and when it was eventually overtaken by sea-level, its growth was terminated, as elsewhere. The topmost date, from a point shortly before this event, is therefore the only one used in drawing the curve.

This is Q663 ; 8970 ± B.P.

7. **Compaction corrections**

It is, as described above, possible to say at what height in relation to the water-table, and therefore to sea-level, the peats were formed. However, before the heights of the peats can be used to determine the absolute height of contemporary sea-level, it is necessary to make allowance for the effects of gravitational compaction. All horizons are now below the height at which they were laid down. Some, resting directly on the solid rock, with little overburden, can have been affected only slightly. Others, with considerable thicknesses of sediments above and below them may have originally been considerably above their present height.

The precision with which the effects of compaction can be estimated varies with the type of sediment. The behaviour of clays under applied overburden pressures is fairly well understood. Peats, however, are of such varied types and undergo such complicated changes during humification, that figures can be only approximate. The height of each dated peat before compaction has been calculated, following the method of Skempton (1970). These values are shown in Table L Also shown is the height of the contemporaneous "effective water-table."

These dates and heights can now be used to construct a sea-level curve (Figure 6). It is shown here compared with various other published curves. It can be seen that there is remarkably good agreement with that of Mörner. In fact it coincides very closely with a lower bounding curve of Mörner's oscillations. This may be explained by the fact that the present curve represents the normal coastal water-table, whilst the high peaks of the Mörner curve may, as previously suggested, be due to the effects of storm

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surges, or similar factors. This agreement with Mörner, since he claims to have eliminated isostatic effects, supports the original assumption of stability in the Bridgwater Bay area.

Jelgersma's curve, with directly-plotted figures from an area which has undergone long-term subsistence, is, as expected, consistently below the curve now presented for the Bristol Channel area.

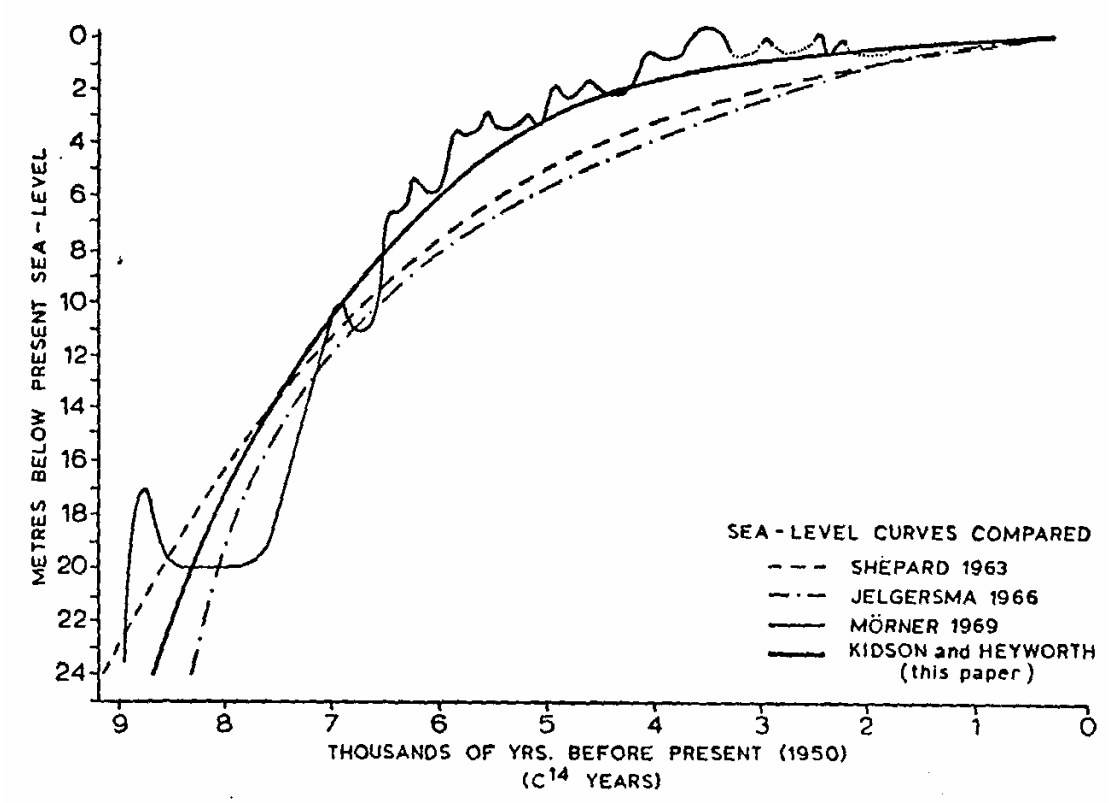

FIGURE 6. Flandrian sea level curves compared.

Shepard's curve is based on world data, Jelsergma's on Netherlands data, Morner's on Southern Swedish data and Kidson and Heyworth's on Bristol Channel material.

8. Age Correction

The curve given in Figure 6 has been further corrected by modifying the C14 dates in accordance with dendrochronological evidence (Suess 1970 and Ferguson 1970). This, whilst introducing additional uncertainties, probably gives a truer picture of the actual rate of change of sea-level rise. This fully corrected curve is shown together with a curve drawn from the completely uncorrected data (Figure 7). The dates are numbered as in Table 1.

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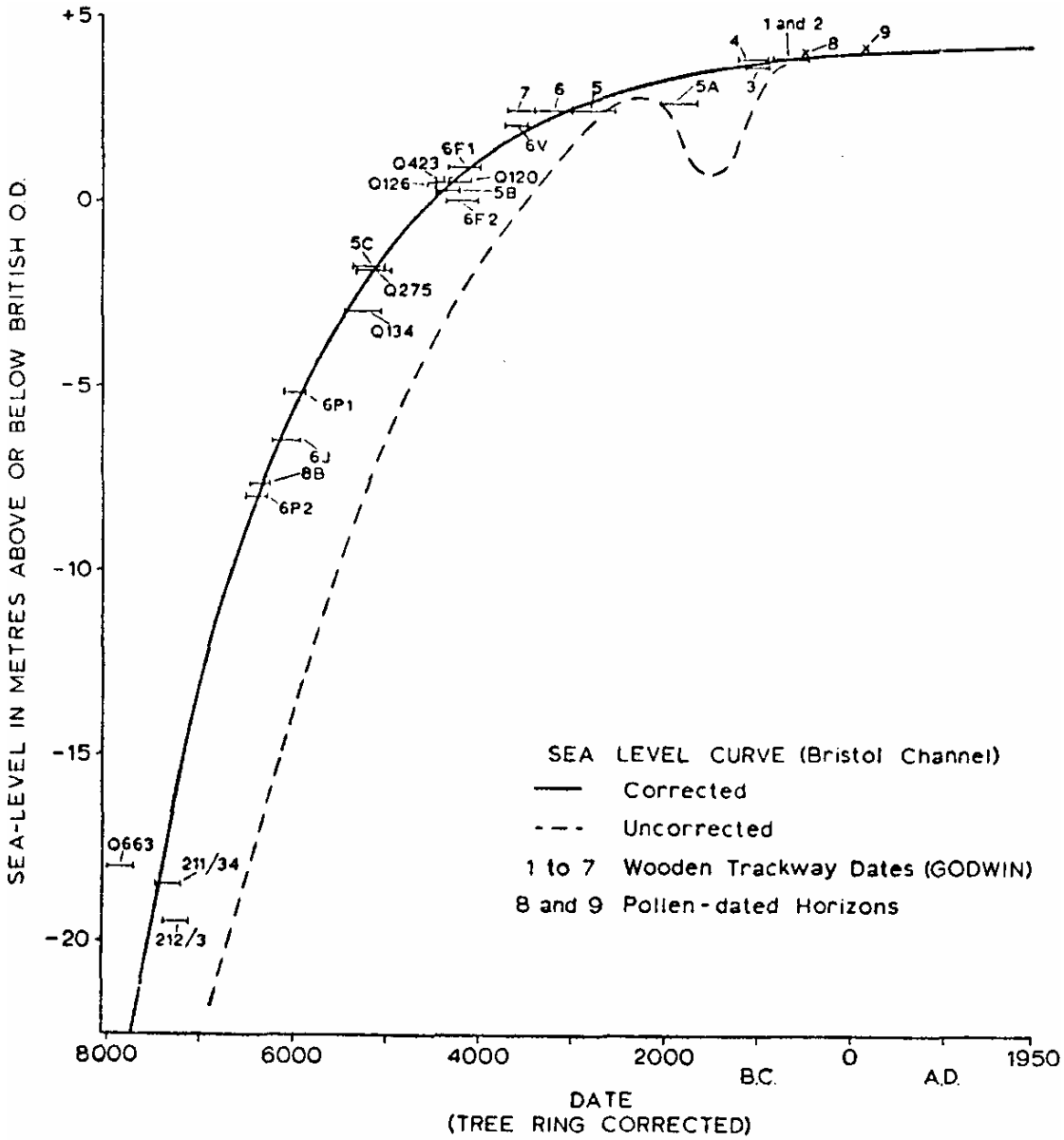

FIGURE 7. Flandrian sea level rise in the Bristol Channel corrected and uncorrected curves.

For explanation of symbols refer to Table I.

For explanation of "sea level" see text.

Pollen dated horizons 8 and 9 are as follows:-

8= "Grenzhorizont" in borehole 6N

9 = Upper flooding horizon at Shapwick Heath (Godwin 1948. p.283).

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In addition to those already discussed, the following "sea-level" peat dates are included, after appropriate correction:

Burnham-on-Sea	Q134	(Godwin, Suggage & Willis 1958)
Shapwick Heath	Q423	(Godwin & Willis 1961)
Telham Moor	Q126	(Godwin & Willis 1961)

It may be significant that the corrected curve is a true exponential one. It can be seen that the rise of sea-level in the last 6000 years has been small. Nevertheless, the importance of this small rise in shaping the present landscape in the area is such that it merits consideration in more detail.

9. Formation of "O.D. clay"

The existence of such an extensive clay surface at a very consistent height suggests a pause in sea-level rise at this point. However, this can not be represented in the curve drawn from other data. The conclusion must be that it was at this height that the rate of rise became so slow that a coastal clay belt could be established, protecting extensive areas inland from continuous flooding by salt-water. At this period peat growth was initiated on this inland clay surface, and once this had occurred, the rate of upward growth is seen to have been so rapid that it exceeded the rate of sea-level rise with the result that it was not overtaken again by the sea until another 4000 years had elapsed. The area of peat in the Somerset Levels is seen from borehole evidence to have reached its maximum extent around the time of the Climatic Optimum. After this date, as the rate of peat growth slowed, the area was progressively reduced. As the rise in sea-level in the last 6000 years is shown to be, at most, *c*.3.5 m, a rate below that expected for peat growth, the overall situation could have changed only very gradually.

10. Romano-British Transgression

There is considerable evidence in the Somerset Levels and in many other areas (Godwin 1943) of a "Romano-British Transgression," when peats which had enjoyed long periods of freedom from the sea were finally covered by estuarine clays. As neither the sea-level curve here presented, nor those of other authors show any recognisable increase in the rate of rise at this date, factors other than a eustatic event must be responsible. This transgression

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is, plainly, the final stages of the gradual overhauling of the peats by sea-level, which had been going on since the Climatic Optimum. The effect of a separate transgression apparently arises from the fact that it was in Romano-British times that the very gradual process reached its maximum extent, and the landward edge of the coastal clay belt reached the main raised-bog areas. Had the process been allowed to continue, it is possible that this effect would be less sharply defined, but it seems that the "Romano-British Transgression" was halted artificially by the beginning of embanking.

11. Conclusion

The main points arising from this investigation are as follows.

Firstly, what appears at first to be a very complex arrangement of peats can be resolved into a simple pattern. This consists of a sloping basal submerged forest peat, buried by marine and estuarine clays which terminate in the generally occurring "O.D. clay" surface. On this clay surface are developed horizontal peats which in places are interrupted by further brackish or fresh-water clays, on which higher horizontal peats are developed. In the thick estuarine clays, purely local horizontal peats are sometimes developed not as a result of any widespread eustatic effect, but as part of the normal processes of coastal change. In other places the basal submerged forest peat occurs at O.D. or above, and so merges with the horizontal upper peats. There is no evidence which necessitates the introduction of oscillations in the sea-level rise to explain these stratigraphical changes.

Secondly, whilst compaction effects are usually small, this is not always so, and they cannot be ignored.

Thirdly, the nature of the peats and clays shows that, throughout the Flandrian, sedimentation has kept pace with sea-level rise over almost the whole of the area. A very high sediment load must have been present to bring about this infilling of the buried river valleys of the Somerset Levels. In the absence of this factor the valleys would now be drowned. Because of the continuous nature of the infilling of the Pleistocene topography,

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the drainage base-level of all the areas of peat-formation has, throughout, been determined by sea-level, so that almost all peats in the Somerset Levels are direct indicators of the corresponding water-table.

Finally, previous assumptions about the isostatically stable nature of the area appear to be confirmed. It should be possible, with confidence, to apply the figures for sea-level rise to other areas of uplift or depression.

ACKNOWLEDGEMENTS. The authors wish to record their thanks to the South Western Road Construction Unit of the Ministry of Transport (Department of the Environment), Soil Mechanics Ltd., and the Somerset River Authority for permission to use bore-hole logs and samples. They are indebted to Miss H. J. Walbancke for valuable assistance with soil compaction problems. To Mr. D. J. Unwin for help with the writing of computer programmes for soil compaction calculations and to Mr. M. M. M. Manton for assistance with photogrammetry and air photo interpretation. They wish particularly to record their thanks to Mr. E. L. Kelting, formerly Chief Engineer of the Somerset River Authority for his valued co-operation throughout the whole period of study.

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THE CENTRAL SOMERSET BASIN

by A. Whittaker

Abstract. The Central Somerset Basin is not only a structural syncline but was also a sedimentary basin into which many of the main Mesozoic stratal divisions thicken from the margins to the centre. The structure is relatively steep-sided and flat-bottomed, suggesting that it may have been initiated as a graben in the Palaeozoic rocks. It is postulated that the bounding faults of the graben may have been active throughout early Mesozoic times.

1. Introduction

In central Somerset, Triassic and Jurassic deposits are disposed in a major, roughly ESE-trending synclinal structure (Fig. 1). Framing the syncline to the north are the Mendip Hills and to the south the Quantock-Cannington Palaeozoic massif. The eastern margin of the basin is obscure but the main structure probably passes beneath Glastonbury to join the Wessex Basin to the east. Westwards, the basin is probably continuous with the Bristol Channel syncline and its *en échelon* folds, in which submarine Triassic and Jurassic rocks have been proved from Bridgwater Bay to farther west than the longitude of Ilfracombe. O. T. Jones (1930) likewise regarded the Bristol Channel downfold as a westward continuation of the Glastonbury Syncline, but he believed that the structure came into existence as a definite basin by Miocene folding. More recently, Owen (1971) accepted the general validity of Jones' views but suggested that the structural history of the Bristol Channel was much more complex and that fracturing on an important scale had occurred. Recent work suggests that the Central Somerset Basin is not only a structural syncline but was also a depositional basin into which many of the Mesozoic stratal divisions thicken from the margins to the centre. Although much detail is known about the Palaeozoic rocks which form the framework of the basin, those forming the floor of its interior have never been penetrated, notwithstanding the fact that boreholes have been sunk to depths of between 600 and 1200 metres at Puriton (McMurtrie 1911) and Burton Row (Whittaker 1972 b) respectively.

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The solid geology of the Central Somerset Basin is masked by the thick alluvial deposits of the Somerset Levels, but solid rocks appear through the drift in several places. The Polden Hills are a low ESE-trending ridge above the level of the alluvial deposits and they divide a belt of Lias sediments to the north from Keuper Marl to the south. The sub-alluvial deposits are known quite well because numerous, usually shallow, boreholes have penetrated them. North of the Polden Hills the sub-drift geological structure has been elucidated by examination of fossil material obtained in boring ; this provides the basis for the structural interpretation of the main on-shore parts of these low-lying drift-covered areas.

This paper will be concerned mainly with the central and southern parts of the basin, as its north-eastern sector was described in some detail by Green and Welch (1965), who noted that the basin had developed progressively throughout Mesozoic times.

2. The pre-Permian topography

The long period of sedimentation which resulted in the accumulation of great thicknesses of Palaeozoic rocks in this region was followed by the Armorican Orogeny. These earth-movements produced great folds and faults in the Palaeozoic strata and gave rise to a nearly E-W structural grain in the country rock. Intense erosion of the uplifted areas followed, so that many hundred metres of Palaeozoic rocks were removed in New Red Sandstone times. The New Red rocks rest unconformably upon the older strata ; in the lower parts of the sequence are arenaceous and rudaceous sediments, commonly largely composed of locally derived clasts. Features such as these allow the inference to be drawn that topography was rugged and that sedimentation took place in intermontane basins in which the products of erosion accumulated. That this was the case in the Central Somerset Basin is suggested by the presence of early New Red Sandstone (? Permian) sediments in both of the deep boreholes (Puriton and Burton Row) which have been drilled within the confines of the basin. Equivalent beds are not known from the swells adjacent to the basin and it is presumed that there is overlap of these strata by younger sediments towards the basin's margin.

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3. The Mesozoic downwarping

Although the New Red Sandstone strata are notoriously difficult to correlate, the base of a group of arenaceous and rudaceous beds is recognisable over much of the area under discussion. This horizon is the base of the Pebble Beds of Ussher ; on a lithostratigraphical basis it is commonly considered to represent the Permian-Triassic boundary. The group is divisible into two parts, a lower rudaceous division, the Pebble Beds, and an upper arenaceous division, the Upper Sandstones. East of the Quantocks the Upper Sandstones are the lowest division of the New Red Sandstone to crop at the surface although conglomerates have been proved underlying the sandstones in the Charlinch area. This demonstrates that at least locally, and possibly over a much wider area, the Upper Sandstones overlap the conglomerates and that the area of sedimentation was being extended as time passed. Upper Sandstones fringe the south-eastern margin of the Quantock Hills and locally contain abundant angular fragments of Devonian material demonstrating that the Quantock massif was exposed and undergoing active erosion during these times. The Upper Sandstones are also exposed, quite extensively, in the area between Cannington and Bridgwater and yet only 2-½ miles (4 km) to the NE their top was proved at a depth of 388 m at Puriton (McMurtrie 1911). Seven miles farther north at Brent Knoll the base of the Pebble Beds - Upper Sandstone division was proved at a depth of 953.44 m; the combined thickness of the two groups increases from about 30 m over the Quantock-Cannington Swell to 60 m in the Central Somerset Basin, suggesting that there was differential subsidence of the area during early Triassic times.

North of a line joining Dodington and Cannington Park there is no trace of the Upper Sandstones at the surface and the succeeding Keuper Marl overlaps earlier Triassic deposits and comes to rest directly on Palaeozoic rocks. Recent work in west Somerset suggests that the Quantock Hills and the Palaeozoic inliers to their north and east formed an upland area which was not largely buried until late in Keuper times. Dips in the Keuper strata of the Cannington area allow an approximate thickness of 150 m to be calculated for the Keuper Marl at the basin's southern margin and this is confirmed from an isopachyte map constructed independently of this information. The isopachyte

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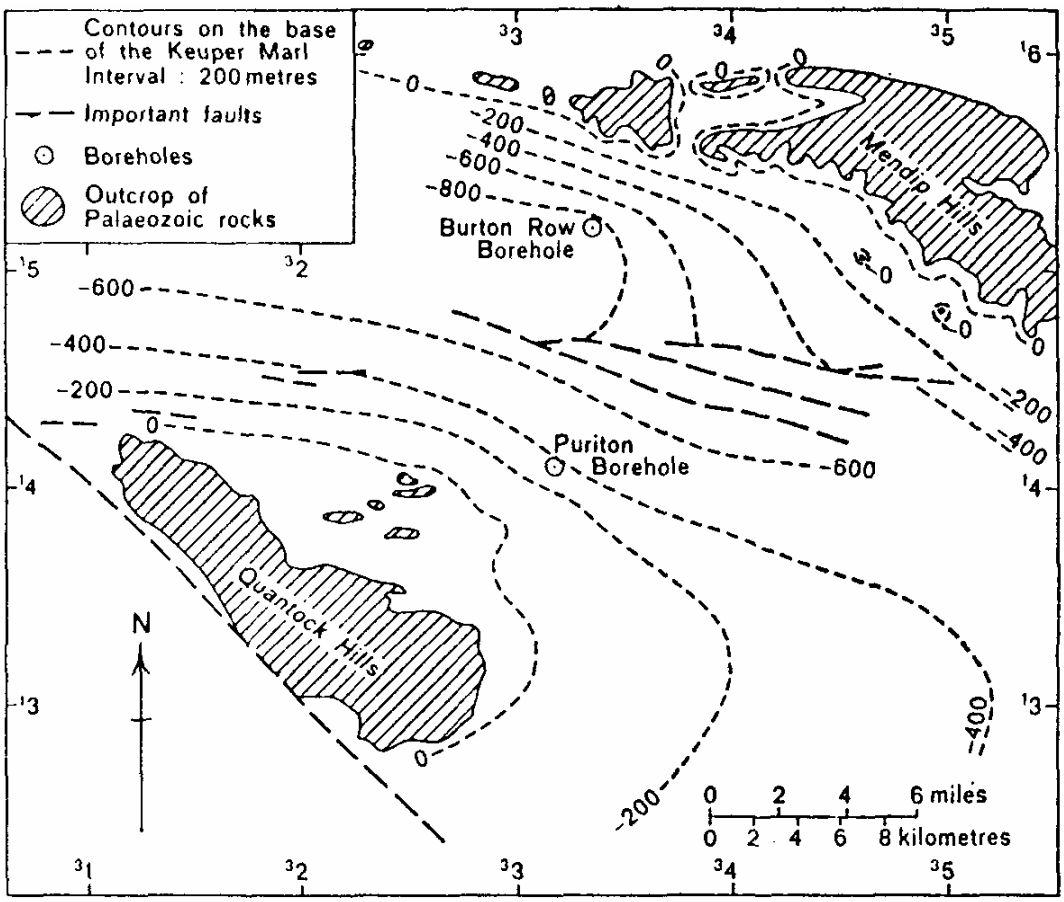

Fig. 1 Structure contour map of the Central Somerset Basin

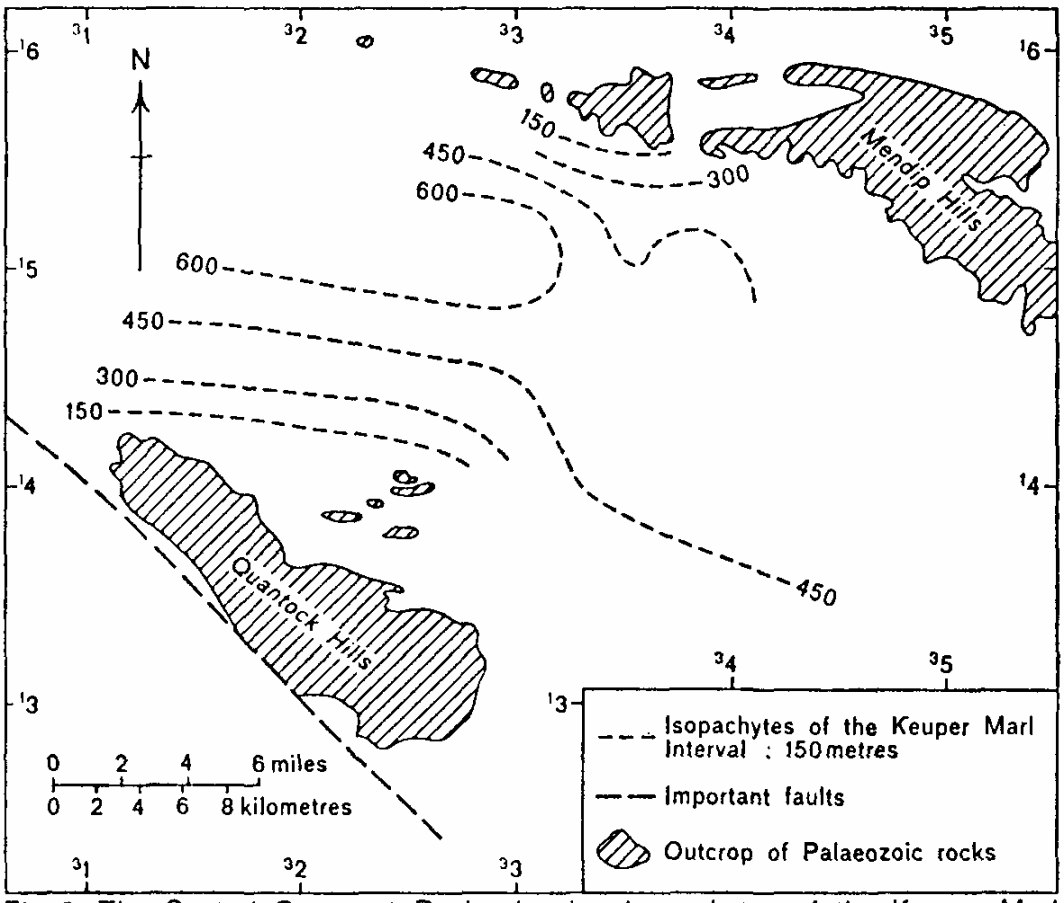

Fig. 2 The Central Somerset Basin showing isopachytes of the Keuper Marl

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map suggests that the Keuper Marl may thicken fourfold into the Central Somerset Basin from its southern margin (Fig. 2). Thus not only was the area of sedimentation being extended in Keuper Marl times but subsidence of the basin area was proceeding at a faster rate.

One of the characteristics of the Permo-Triassic major depositional basins is the occurrence of thick developments of rock salt. The Central Somerset Basin is no exception ; rock salt was proved at both Puriton and Burton Row. Stratum contours on the top surface of the bedded rock salt show a broad synclinal structure, and isopachytes demonstrate that the saliferous beds increase in thickness from 30 m near the margin of the basin to perhaps 90 m near its centre. It is possible that the salt beds `wedge out' at the basin's margins to be overlapped by higher Keuper Marl, but at the southern margin it is also possible that the saliferous strata ‘crop’ beneath the alluvium. Because of ground-water circulation, saliferous beds cannot reach the surface in this country and their position is usually marked by masses of collapsed and brecciated marls. Boreholes south of the Polden Hills show that the sub-drift valley of the River Parrett follows a course different from its present position. Soft, disturbed ground such as one may expect at the ‘crop’ of saliferous marls would make an ideal and preferential channel for a river and the old valley seems to parallel the strike and follow what could be the ‘crop’ of the saliferous marls. In connection with this, a borehole near Bridgwater penetrated the drift-solid interface and proved brackish water or brine. Another clue to the presence of brine may be found in the Burrow Bridge area where there is a locality known as Salt Moor.

The Tea Green Marl and Grey Marl, considered as one unit, thicken from the Puriton area (24 m) basinwards to Brent Knoll (38 m).

Of the Jurassic, only the Blue Lias (*Psiloceras planorbis* to *Arnioceras semicostatum* zones) is exposed in the west Somerset coast sections and it is about 122 m thick (Whittaker 1971). The Burton Row Borehole Blue Lias and that of the Polden Hills are very similar in detail to the coastal exposures. Within the basin the Lias stratigraphy is remarkably consistent laterally ;

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only in the vicinity of the swells are changes evident. At Cannard's Grave near Shepton Mallet, the Blue Lias is relatively very attenuated, achieving a thickness of some 7.5 m (Donovan 1958) compared with the 122 m recorded farther west. Facies changes are evident in the Lower Lias over the Mendip Swell (Green and Welch 1965). In regard to the southern margin of the basin, unfortunately the Lias escarpment parallels the edge, and is some distance to the north of the Quantock Swell. However the crop of the Lower Lias crosses the sub-Mesozoic continuation of the swell in the vicinity of Somerton and the strata of the *P. planorbis* Zone thin markedly, but maintain the same facies, from Cossington in the Polden Hills to the Street area. The only other place where Lias beds approach the swell is at Blue Ben, where inliers of Hangman Grit, mantled by Keuper Marl, are close to the coast. Hereabouts the Lias is faulted into contact with Keuper Marl, the fault throwing : down some 167 m to the north, and the *Schlotheimia angulata* Zone is 31 m thick compared with between 40 m and 52 m elsewhere in the basin. Again there is no apparent facies change in these Blue Lias beds. Although the Lias is at sea level near St. Audrie's there are faults affecting these beds with throws of 200 m in the immediate neighbourhood and it is quite conceivable that at one time thin Lias deposits were present over the Quantock Hills.

The thickness of Lias at Brent Knoll is of the order of 538 m, the combined Middle and Upper Lias representing about 164 m of this total. Basinal thicknesses for the combined Middle and Upper Lias (l40 m) are maintained in the Glastonbury area but these strata are very thin or absent in the main escarpment outside the basin.

Thus the history of Liassic sedimentation is in general one of subsidence and sedimentation over much of the area but with much greater downwarping of the basin to allow accumulation of considerable thicknesses of sediment.

4. The post-Liassic deformation

The Rhaetic forms a good datum for structural analysis as this formation is readily recognisable, and where present is usually in the same facies over much of the area. Rhaetic deposits occur at a height of +244 m O.D. over the Mendip Swell, at +61m O.D. in the Polden Hills near the Quantock Swell, and yet are

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depressed to -411 m at Brent Knoll ; since the sediments were formed under similar conditions in a similar environment it follows that a relative downwarping of 650 m has taken place since Rhaetic times. Some of this downwarping undoubtedly took place during Mesozoic times but some also is due probably to post-Mesozoic movements. Small-scale E-W folds are present in the Liassic rocks of the west Somerset coast and within the landward area of the basin. Large-scale faults follow a similar trend and likewise affect Liassic strata. Besides normal faults, some reverse faults and thrusts have been observed in the west Somerset Lias and an important NW-trending post-Liassic trans-current fault is present near Watchet (Whittaker 1972 c). E-W structures affect later Mesozoic strata in the area to the east of that under discussion and it seems probable that some of the basin's structures are post-Mesozoic. It is considered likely that post-Mesozoic movements would accentuate earlier intra-Mesozoic structures as well as initiating new ones.

5. Conclusions

The whole impression of the basin at successive stratigraphic levels as gained from a study of structure contour maps, is that it is a rather steep-sided and relatively flat-bottomed feature. Thus, contours on the present-day sub-Permian surface are closely spaced near the edges of the trough suggesting the possibility of fault scarps ; this suggests in turn that the structure may have been initiated as a graben in the Palaeozoic rocks. Thickness changes seem to occur rapidly in the vicinity of the swells which frame the basin, and differential subsidence of the trough throughout New Red Sandstone and early Jurassic times allows the inference to be drawn that the postulated bounding faults of the basement graben were active, possibly intermittently, throughout the whole of this time. The basin bears close comparison with the Severn Basin (Audley-Charles 1970, Whittaker 1972 a) where graben tectonics have been invoked to explain similar features. The various datum levels used for contouring (base Keuper Marl, base Lias) show that the structure contours near the swells become less closely spaced at successively higher levels in the sequence, suggesting that the structures at the basin's margins become less intense upwards through the succession.

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In conclusion it is clear that the basin is not due to Tertiary movements alone, but is the result of a long history of down-warping which was initiated at the close of the Armorican Orogeny and which developed throughout Mesozoic and later times.

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AMMONITE FAUNAS OF THE UPPERMOST KIMMERIDGE CLAY, THE PORTLAND SAND AND THE PORTLAND STONE OF DORSET

by J. C. W. Cope and W. A. Wimbledon

Abstract. The succession of ammonite faunas from the uppermost part of the Kimmeridge Clay and the Portland Beds is described. On the basis of work so far completed new correlations within Dorset are suggested.

In this account the Kimmeridge Clay and Portland Sand ammonite faunas are reported on by J. C. W. Cope and those of the Portland Stone by W. A. Wimbledon. The results of collecting from the top part of the Kimmeridge Clay will shortly be published, together with descriptions of new ammonite faunas and a new zonal scheme. Work on the Portland Beds is not so far advanced, and the description herein may be considered a progress report on results so far obtained.

The ammonite faunas of the Upper Kimmeridge Clay to a position midway up the Pectinatus Zone have already been described (Cope 1967). The series of clays above this pass with gradual lithological change into the Portland Sands above.

Ammonites showing a progressive series of changes have been collected from the upper part of the Pectinatus Zone. Dimorphism is well displayed in species of *Pectinatites,* and in the upper part of the Pectinatus Zone the horn of the microconch specimens becomes progressively smaller. Associated with this is a progressive coarsening of the ribs on the outer whorl. The genus *Pectinatites* becomes gradually more *Pavlovia*-like, until the horn of the microconch is lost altogether, and the ammonites are true *Pavlovia.* This change from *Pectinatites* to *Pavlovia* shows that the origin of the latter genus differs from that hitherto supposed. Formerly it was believed that the subfamily Dorsoplanitinae evolved from the genus *Subdichotoinoceras* (e.g. Arkell 1957:L332), but now it appears that this subfamily arose from the Virgatosphinctinae.

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The lowest *Pavlovia* faunas do not appear to have been recorded in Britain outside of Dorset, and the species of *Pavlovia* are all undescribed forms. These ammonites are succeeded by the Pallasioides Zone faunas. This zone known for many years from the Hartwell Clay of the Aylesbury district (Neaverson 1925), was believed by Neaverson to lie above his Rotundum Zone. It has been demonstrated in the South Midlands by Casey (1967) that the Pallasioides Zone lies below and not above the Rotunda Zone. Forms characteristic of the Hartwell Clay have now been found in Dorset.

The overlying Rotunda Zone yields abundant crushed ammonites, except those from the Rotunda Nodule Beds which are finely preserved 'Pavlovids in calcareous nodules. A few metres above the latter horizon, ammonites become significantly rarer and new forms appear. From these higher beds, up to the base of the Massive Bed (which marks the base of the Portland Sands in Purbeck) no identifiable ammonites have been obtained in the past. Recently, however, these beds have yielded a series of ammonites which, though rather indifferently preserved, form a readily recognisable faunal assemblage. As well as *Pavlovia* (sensu stricto) these highest beds have yielded a new genus of ammonite retaining pavlovid inner whorls but bearing virgatitid ribbing on the outer whorl.

The Massive Bed at the base of the Portland Sand has yielded *Progalbanites,* an ammonite which also characterises the beds above up to the White Cementstone Band. At about this latter horizon the genus *Epivirgatites* appears and characterises horizons up to the base of the lowest Parallel Band (Arkell 1947). On Portland, *Epivirgatites* occurs in the Black Nore Sandstone suggesting that Arkell's correlation of this horizon with the Massive Bed of Purbeck should be revised. The highest horizons in the Portland Sand (the 'Black Sandstones' of Arkell) which are dolomites, are characterised by large serpenticone perisphinctids similar to *Gyromegalites polygyralis* Buckman, together with species of *Crendonites.* With these two genera there occur other species similar in general aspect to those characterising the Basal Shell Bed of the Portland Stone above.

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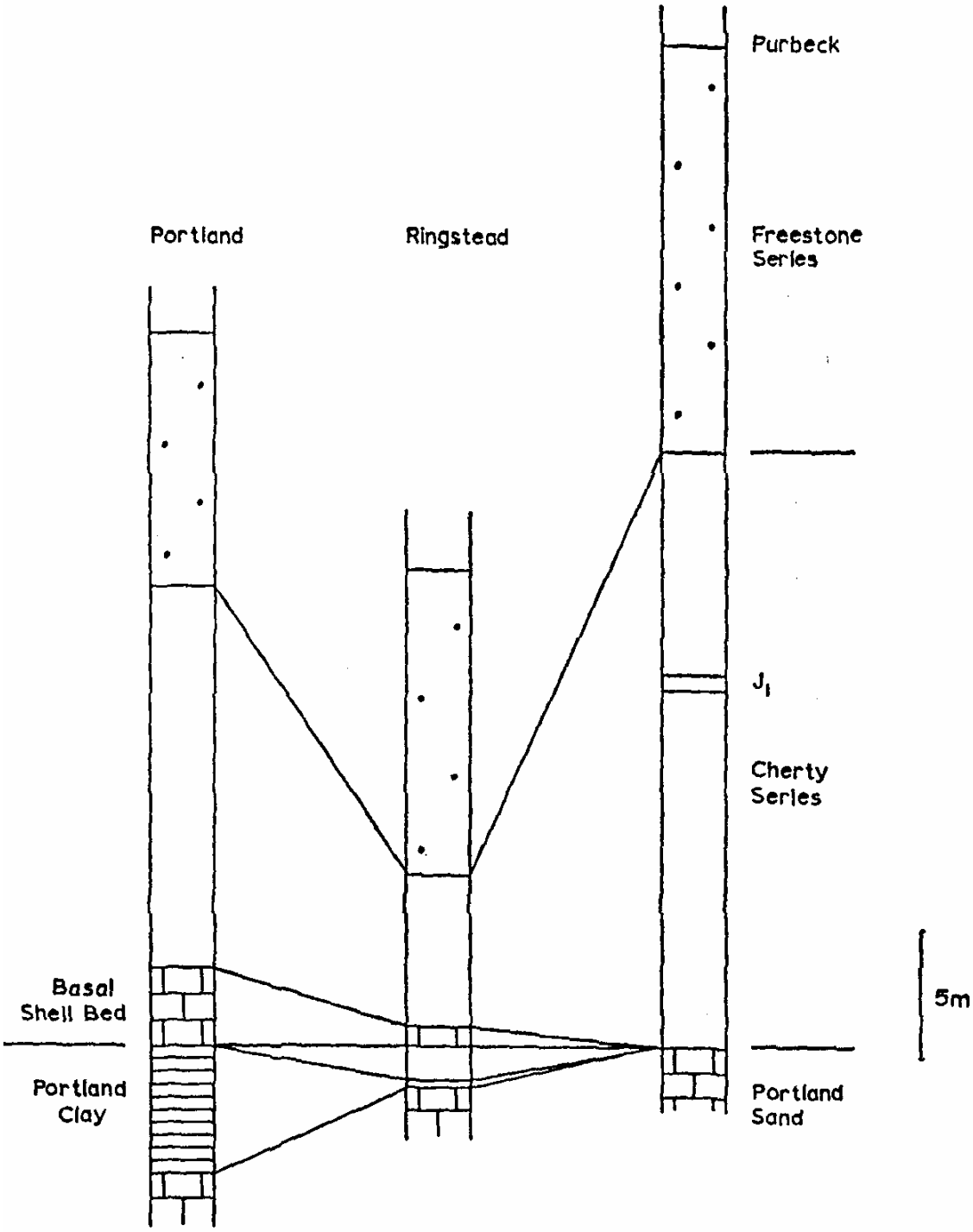

Present correlation between Portland and Purbeck

FIGURE 1.

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Previously, the Portland stone of Dorset has been divided into a Freestone and Cherty Series, the former consisting predominantly of bioclast and ooid sands and the latter of cherty micrites. Correlation between Portland and Purbeck has been attempted along lithological lines, the Freestones of the one area being taken as the equivalent of the other and so on.

Examination of the ammonite faunas has shown that the Cherty Series as a whole, in Portland and Purbeck, are not true equivalents. On Portland, ammonites are not found in the Portland Clay but are abundant in the Basal Shell Bed and lower part of the Cherty Series. In the Isle of Purbeck the lower part of the Cherty Series is barren except for trace fossils. It is about the horizon known as the Puffin Ledge, Arkell's bed J, (Arkell 1935), that ammonites are common (i.e. in the upper part of the Cherty Series).

Triplicate-ribbed inner whorls of the larger forms occur in the Basal Shell Bed of the western exposures. These inner whorls include such forms as *Ammonites triplicatus* Blake and *Kerberites portlandensis* Cox. The mature individuals seemingly are not members of the *Titanites* group but include forms figured from Buckinghamshire by Buckman.

One form which can be used for correlative purposes is related to *Crendonites leptolobatus* Buckman. This is an evolute form of about 150mm diameter with bifurcating ribbing and finely ribbed inner whorls. It has a distinctive collared mouth border. The species occurs in the Basal Shell Bed of Portland and the western mainland exposures, in the Isle of Purbeck it is found in the upper part of the Cherty Series. In addition to this evidence, the ‘gigantids’ which also are present lend support to the suggested equivalence of the Basal Shell Bed and Arkell's bed J_1. Using this scheme it would seem that the Portland Clay is equivalent to all or part of the Lower Cherty Series of Purbeck.

The well known ammonites of the Freestone Series tend to confirm the approximate time equivalence of the Freestone sediments at the two ends of the outcrop. The earliest members of the *Titanites* ‘group’ so far found are forms from the lower Cherty Series of Portland. These have bifurcating ribs together with the typical common single rib so characteristic of *Titanites*.

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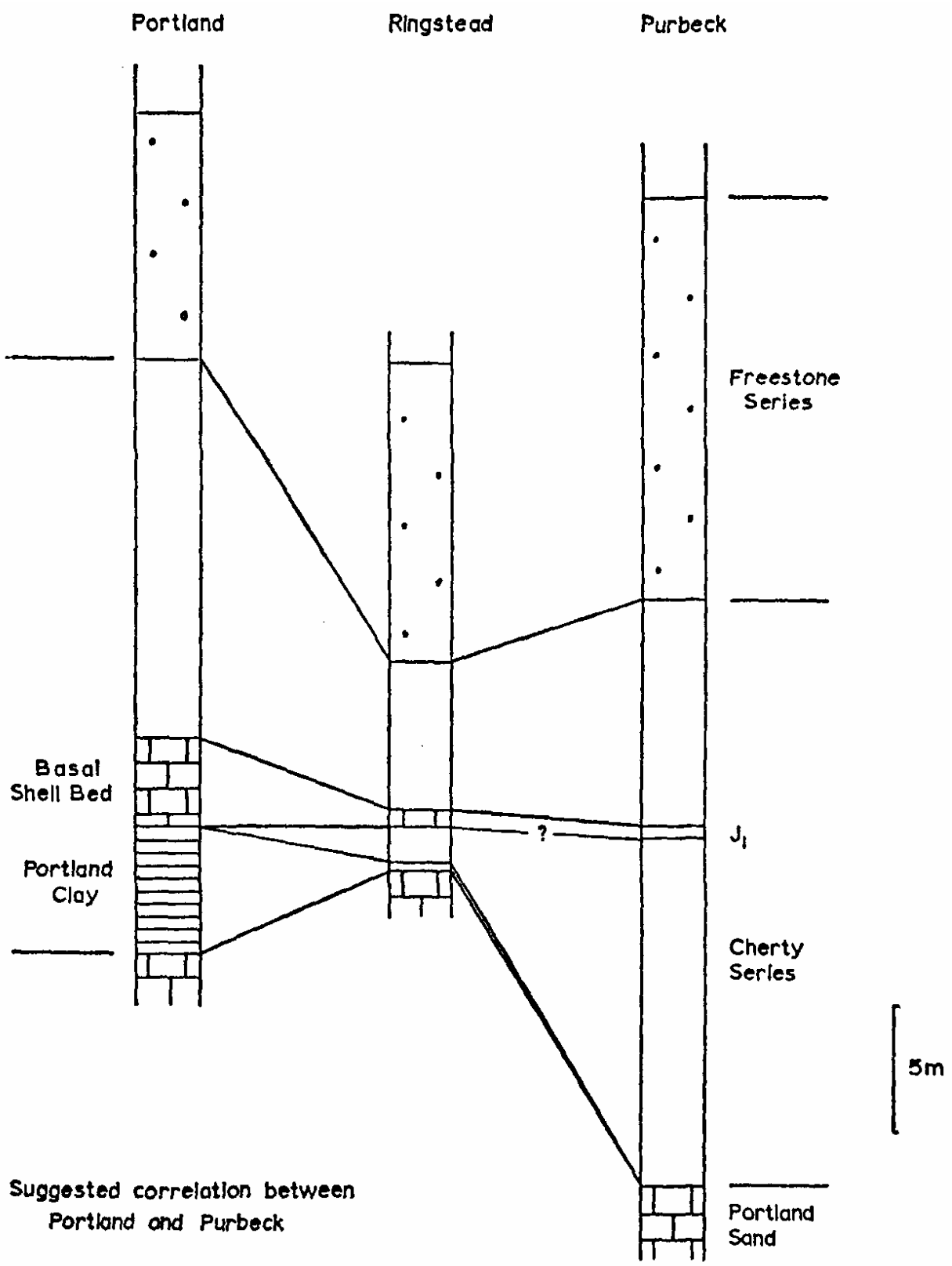

Suggested correlation between Portland and Purbeck

FIGURE 2.

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SOME OBSERVATIONS ON THE CHERT BEDS (UPPER GREENSAND) OF SOUTH WEST ENGLAND

by M. B. Hart

Abstract. The Chert Beds (Upper Greensand) in South West England are shown to be of Lower Cenomanian age. The stratigraphy and depositional history of the Chert Beds is discussed in the light of new palaeontological evidence. Recent analyses of Upper Greensand fossils and glauconite grains suggest that there was some important early diagenetic redistribution which may have a bearing on more general theories of chert formation.

1. Introduction

The Upper Greensand of South-west England is poorly known and its stratigraphic position is disputed. It has also been involved in the controversy about the formation of cherts (Tresise 1960, 1961). This account is primarily intended as a stratigraphic analysis of the Upper Greensand in general and of the Chert Beds in particular.

2. **Dating of the Upper Greensand**

Although several workers have discussed the stratigraphy of the Upper Greensand (for a detailed bibliography see Hancock 1969 ; also Drummond 1970, Kennedy 1970) most have relied on the palaeontological evidence of Spath (1926). Only Wright (in Arkell 1947) and Kennedy (1970) have provided new macrofaunal evidence. Hart (1971) presented the first micropalaeontological evidence for the age of the Upper Greensand in the South-west of England. That account is amplified here, and comparison made with the ammonite chronology.

The ammonite evidence from the Upper Greensand is imperfect and many problems stem from the correlation of this poorly known sequence with the detailed faunal succession of the Gault Clay of South-east England. Hancock (1969) summarised the zonal determinations of the Upper Greensand, and some of these conclusions are discussed in this account. The Chert Beds have never been properly defined as a stratigraphical unit and

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there is no accepted type locality. It is presently proposed that the Chert Beds be regarded as a member of the Upper Greensand Formation. The main guide to the stratigraphic position of the Chert Beds comes from central Dorset (Fig. 1) where the sands with cherts appear in the succession at Maiden Newton (Standers Mill Plantation ; SY587976). Passing westwards from the area of the Mid-Dorset Swell (Drummond 1970, Kennedy 1970) the first cherts appear in the coarse sands which lie stratigraphically between the '*Exogyra* Sandstone' and the 'Eggardon Grit.'

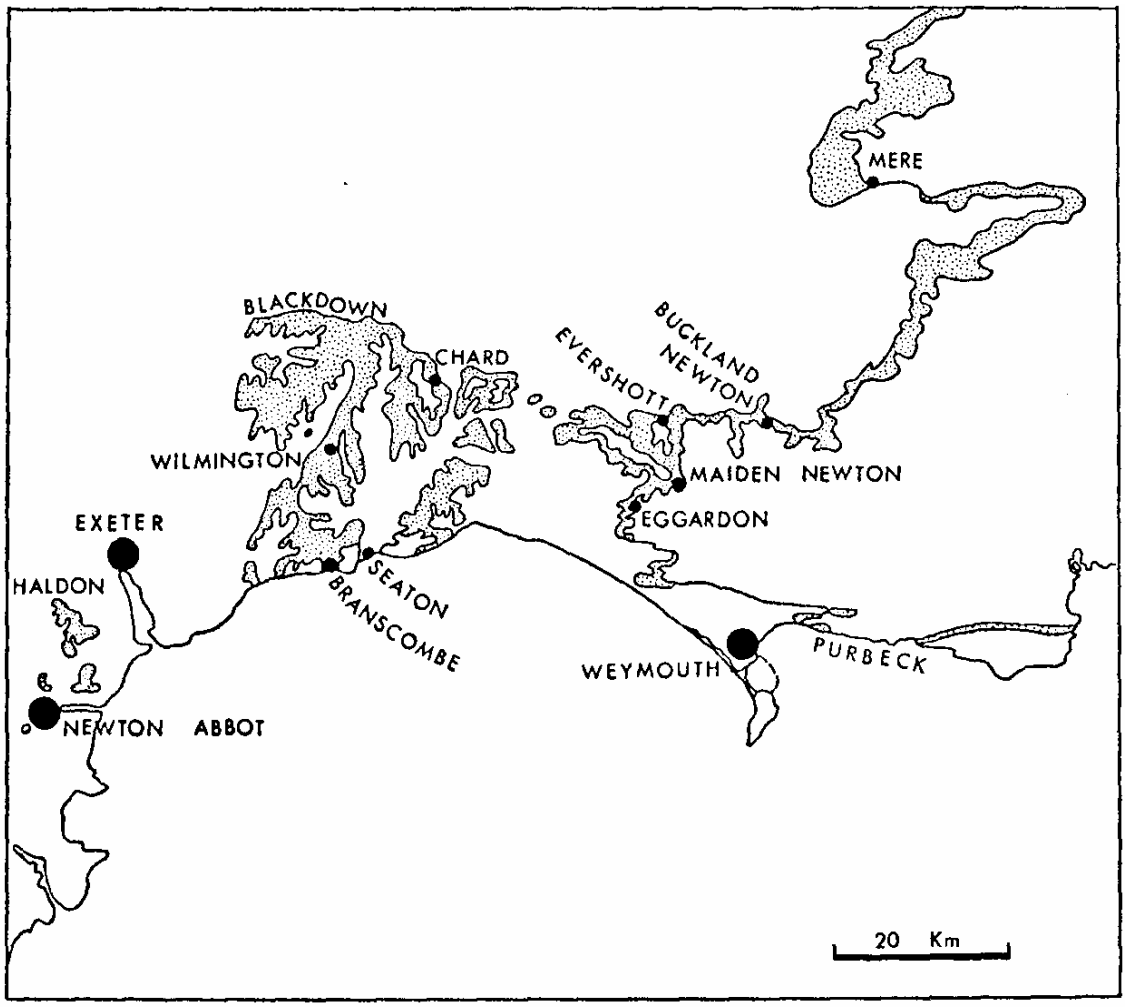

FIGURE 1.

These two lithostratigraphic units can be dated satisfactorily from both microfaunal and macrofaunal evidence. The '*Exogyra* Sandstone' is a glauconitic, calcareous sandstone, that contains abundant *Exogyra obliquata* (Pulteny), and other silicified bivalves.

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The occurrence of *M.* (*Mortoniceras*) aff. *commune* Spath places this sandstone within the *Callihoplites auritus* or *Mortoniceras aequatorialis* Subzone of the Upper Albian (Kennedy 1970). The 'Eggardon Grit' was named by Wilson *et al.* (1958) from the type locality on Eggardon Hill, where it had been previously described as the 'calcareous sandstone' by Jukes-Browne and Hill (1900, 1903). It is also known as the 'Top Sandstones' on the Devon coast. The 'Eggardon Grit' lacks glauconite and at its type locality, contains a distinctive, non-phosphatised, Lower Cenomanian ammonite fauna (Kennedy 1970:642). The microfauna of the Eggardon Grit has provided supporting evidence in mid-Dorset and South-east Devon, where the usual microfauna is associated with *Orbitolina lenticularis* (Blumenbach) - see Hart (1971). This microfauna is associated with a level (Zone 9 of the mid-Cretaceous micro-faunal zonation proposed by Carter and Hart, M.S.) high in the Lower Cenomanian.

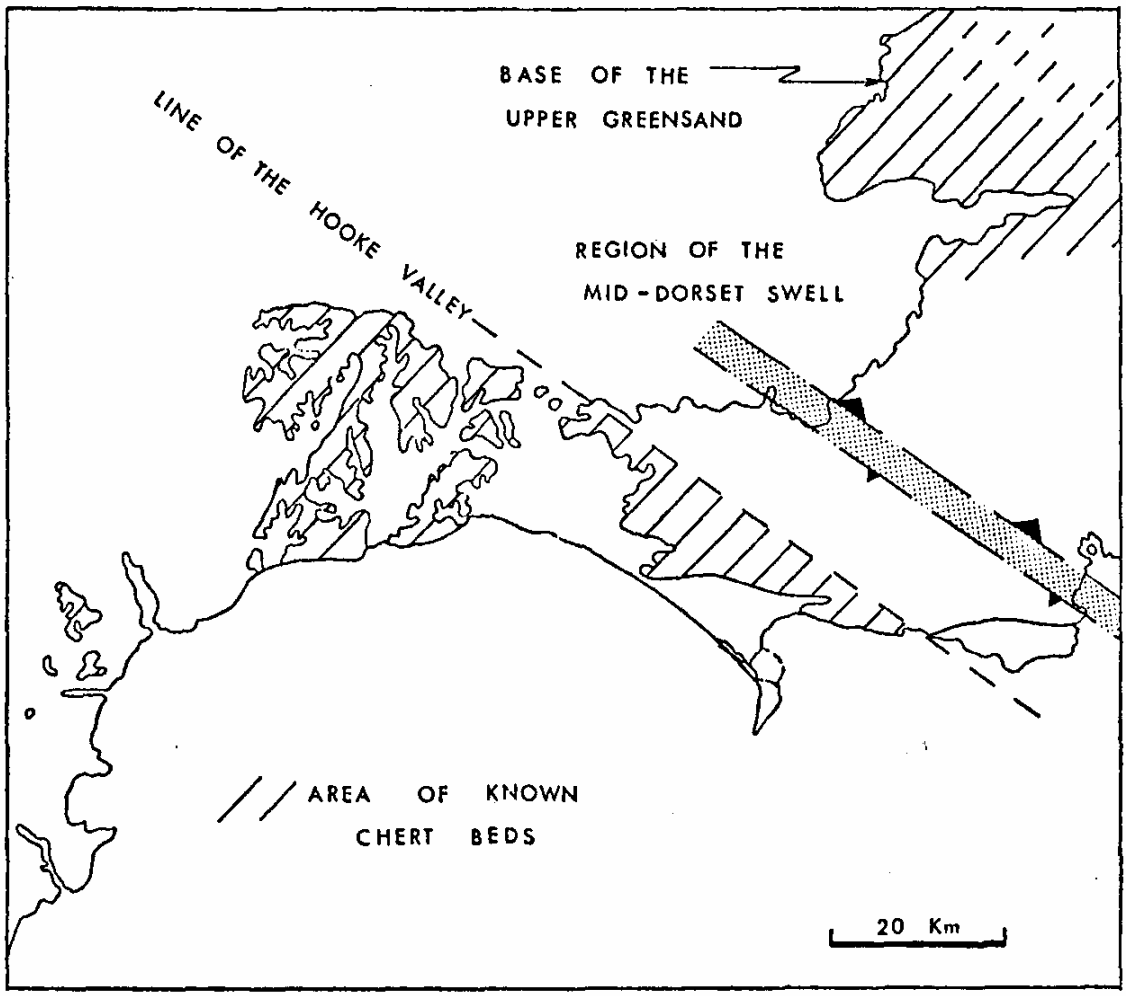

FIGURE 2.

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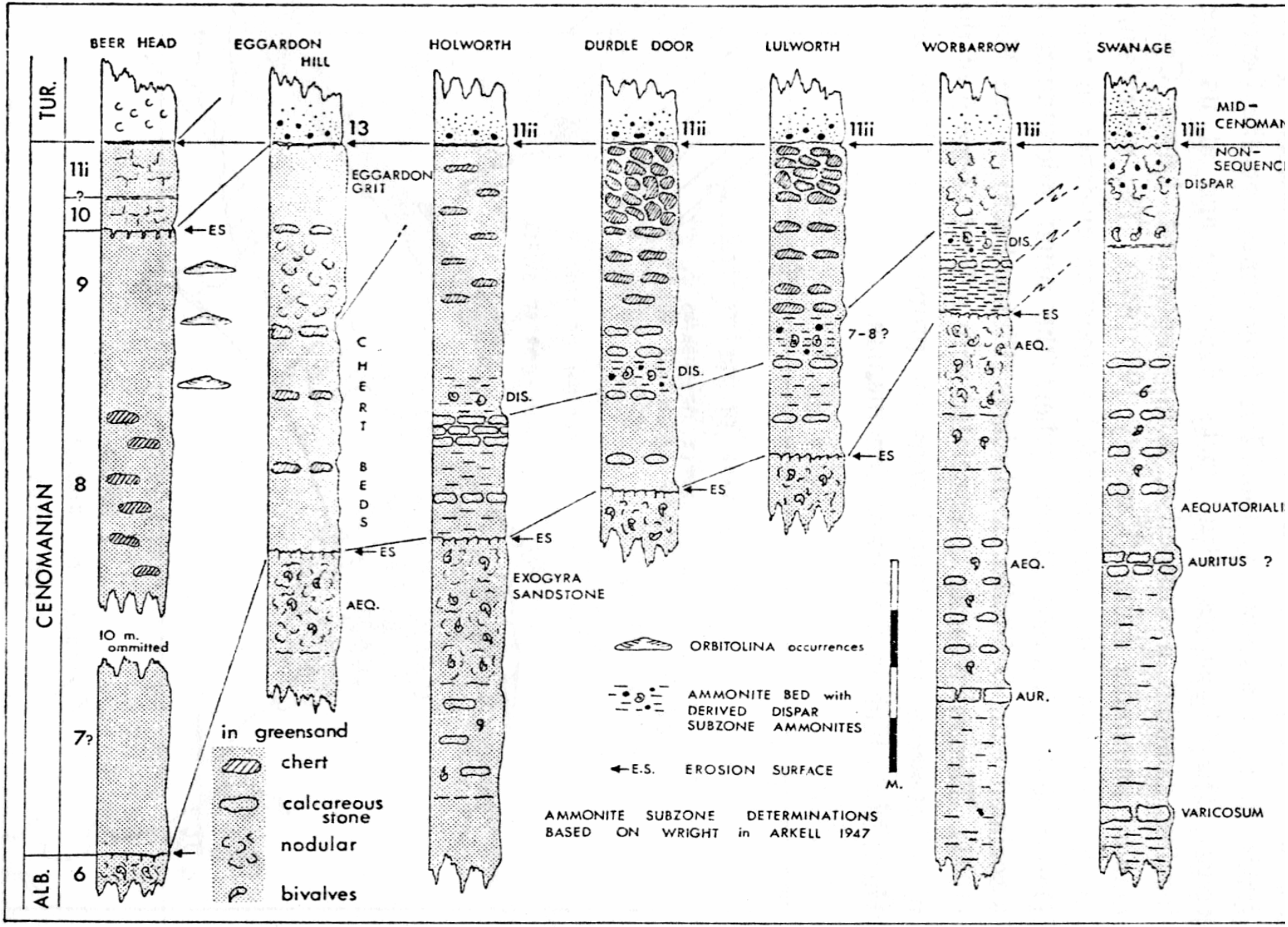
BEER HEAD
EGGARDON HILL
HOLWORTH
DURDLE DOOR
LULWORTH
WORBARROW
SWANAGE
TUR.
CENOMANIAN
ALB.
11i
10
9
8
7?
6
13
11ii
EGGARDON GRIT
CHERT BEDS
ES
AEQ.
DIS.
EXOGYRA SANDSTONE
7-8?
MID-CENOMAN
NON-SEQUENCE
DISPAR
AEQUATORIALIS
AURITUS ?
VARICOSUM
AUR.
10 m. ommitted
in greensand
chert
calcareous stone
nodular
bivalves
ORBITOLINA occurrences
AMMONITE BED with DERIVED DISPAR SUBZONE AMMONITES
E.S. EROSION SURFACE
M.
AMMONITE SUBZONE DETERMINATIONS BASED ON WRIGHT in ARKELL 1947

FIGURE 3.

The '*Exogyra* Sandstone' and 'Eggardon Grit' show a constant relationship to the Chert Beds, which gradually thicken south-westwards from the line of the Hooke Valley (Fig. 2). This feature has been observed by several authors (Wilson *et al.* 1958, Kennedy 1970, Drummond 1970) and a relationship to the Mid-Dorset Swell has also been suggested. This indicates that the Chert Beds at their maximum development, near Beer and Seaton, must occupy an interval spanning either the whole of, or a part of, the *Stoliczkaia dispar* Zone (Upper Albian) and the *Hypoturrilites carcitanensis* assemblage zone (Lower Cenomanian) of Kennedy (1969). The macrofaunal evidence for the accurate placing of the succession within this range is inadequate for it appears to be confined to a single record of *Mortoniceras* gr. *stoliczkaiu* from about this stratigraphic level at an unspecified locality near Charmouth. The localization is quite unsatisfactory for modern biostratigraphical work, but the majority of workers have nevertheless taken this specimen as indicative of a *S. dispar* Zone dating for the whole of the Chert Beds in the South-west of England (eg. Kennedy 1970:642).

On the Purbeck coastline (Fig. 3) the sands with cherts appear below the Chalk and above the lateral equivalent of the '*Exogyra* Sandstone' (the 'Eggardon Grit' appearing in the succession further to the west). This area was discussed in detail by Wright (in Arkell 1947) who recorded the presence of an extensive *S. dispar* Zone fauna (largely belonging to the M. (D.) *perinflatum* Subzone - the *A. substuderi* Subzone apparently being absent). The fauna is totally phosphatised and is in the form of broken, pale buff to cream casts, concentrated in a bed of dark green marly greensand. Wrights' *S. dispar* Zone faunas appear (1947: 185, 186) to come only from the 'Ammonite Bed', which can be traced from Swanage to Holworth House along the coastline. This faunal association was compared, by Wright, to that found in the derived fauna of the Cambridge Greensand. However the microfauna recovered from the sediment in which the *S. dispar* Zone ammonite fauna was collected is almost certainly of Lower Cenomanian age, and compares favourably with the succession at Dover, indicating an *H. carcitanensis* assemblage zone age. It can only be suggested that the ammonite fauna of the Dorset *S. dispar* Zone is derived, and that it lies in the same stratigraphic position as that recorded from Cambridgeshire (Hart 1973). In

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Cambridgeshire the Gault Clay below the Cambridge Greensand is of *C. auritus* and *M. aequatorialis* Subzone age, and this would also appear to be the case in Dorset. The 'Ammonite Bed' can be traced along the Purbeck coastline as far as Holworth House, and although it has not been recorded west of that point, it is possible that the well-quoted ammonite from Charmouth came from this horizon in the lowermost part of the Chert Beds.

There would therefore appear to be little evidence for the existence of *S. dispar* Zone Upper Greensand in Dorset and south east Devon. Over the whole of this area the Chert Beds are underlain by *M. aequatorialis* (?) Subzone '*Exogyra* Sandstone', and overlain by the 'Eggardon Grit' (or 'Top Sandstones') which appears to belong to the Lower Cenomanian. The determination of a Lower Cenomanian age for the Chert Beds is substantiated by the occurrence of *O. lenticularis* (Hart 1971).

There are problems in the dating of the Upper Greensand in the Haldon Hills, which concern a single unlocated specimen that has been determined as the nucleus of a *Stoliczkaia,* and hence of Upper Albian age. Recent work by Dr. R. A. Edwards has shown the presence of highly fossiliferous cherts at Babcombe Copse (SX.869766) and Sands Copse (SX.867757), on the edge of the Bovey Basin. These fossiliferous cherts are not seen *in sitit* in these quarries but they can be placed in the local succession at Smallacombe Goyle (SX924766). These brown fossiliferous cherts contain abundant *O. lenticularis* (the fauna being identical to that found in blocks of Upper Greensand limestone at Wolborough (SX85506995)) and have also yielded three ammonites that have been determined (Wood 1970) as *Mantelliceras* sp., *Hyphoplites* sp. cf. *H. pseudofalcatus* (Semenow), and *Turrilites* sp cf. *acutus* Passey ; they indicate an admixture of Lower and Middle Cenomanian forms. While the first two identifications would agree with the Lower Cenomanian dating of the Chert Beds put forward here, the occurrence of *T. acutus* remains a problem. The *T. acutus* assemblage fauna of Kennedy (1969) always occurs above the mid-Cenomanian non-sequence (Hart 1971) and as this fauna has not been recorded *in situ* west of mid-Dorset, there is clearly a discrepancy that remains to be resolved. As this involves ammonite taxonomy and identification the author cannot comment on this problem at the present time.

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The microfaunal and macrofaunal evidence (apart from the *T. acutus* determination) suggests that the Chert Beds are of Lower Cenomanian age, and that they appear to equate with the *H. carcitanensis* assemblage zone of the South-east of England.

3. The problem of chert formation

There are several important factors to be taken into account when considering the concentration of silica at this level in the succession.

a. While the cherts in the Chert Beds indicate strong silicification or silica concentration, the overlying Cenomanian Sands at Wilmington (SY208999) show evidence of only moderate beekitization. Further, there is no evidence of silicification above the mid-Cenomanian non-sequence (Hart 1971), indicating that either the source of the silica ceased at this level, or that silica enrichment during the interval of the non-sequence was completed before the resumption of sedimentation.

b. The limestones at Wolborough show incipient silicification, which, had it continued, would have produced cherts like those found at Babcombe Copse - the *Orbitolina* fauna from both localities being almost identical.

c. The cherts occur in distinct bands, about 70 cms apart, which appear to be part of a sedimentological cycle. The lowest levels of the Lower Cenomanian over the greater part of southern England are characterised by alternating beds of marly and more calcareous chalk. These cycles are generally the same thickness as those found in the Chert Beds, and as they both seem to be contemporaneous it is probable that they were formed under similar influences. Of particular interest is the abundance of the hexactinellid sponge *Exanthesis labrosus* (T. Smith) in the more calcareous units of the lowermost Lower Chalk. These sponges are greatly reduced in numbers above the *H. carcitanensis* assemblage zone.

Further evidence has been obtained from an X-ray dispersive analyser that has been used in conjunction with a scanning electron microscope. Glauconite grains (which are usually dark green in colour) from within the chert-bearing greensands are

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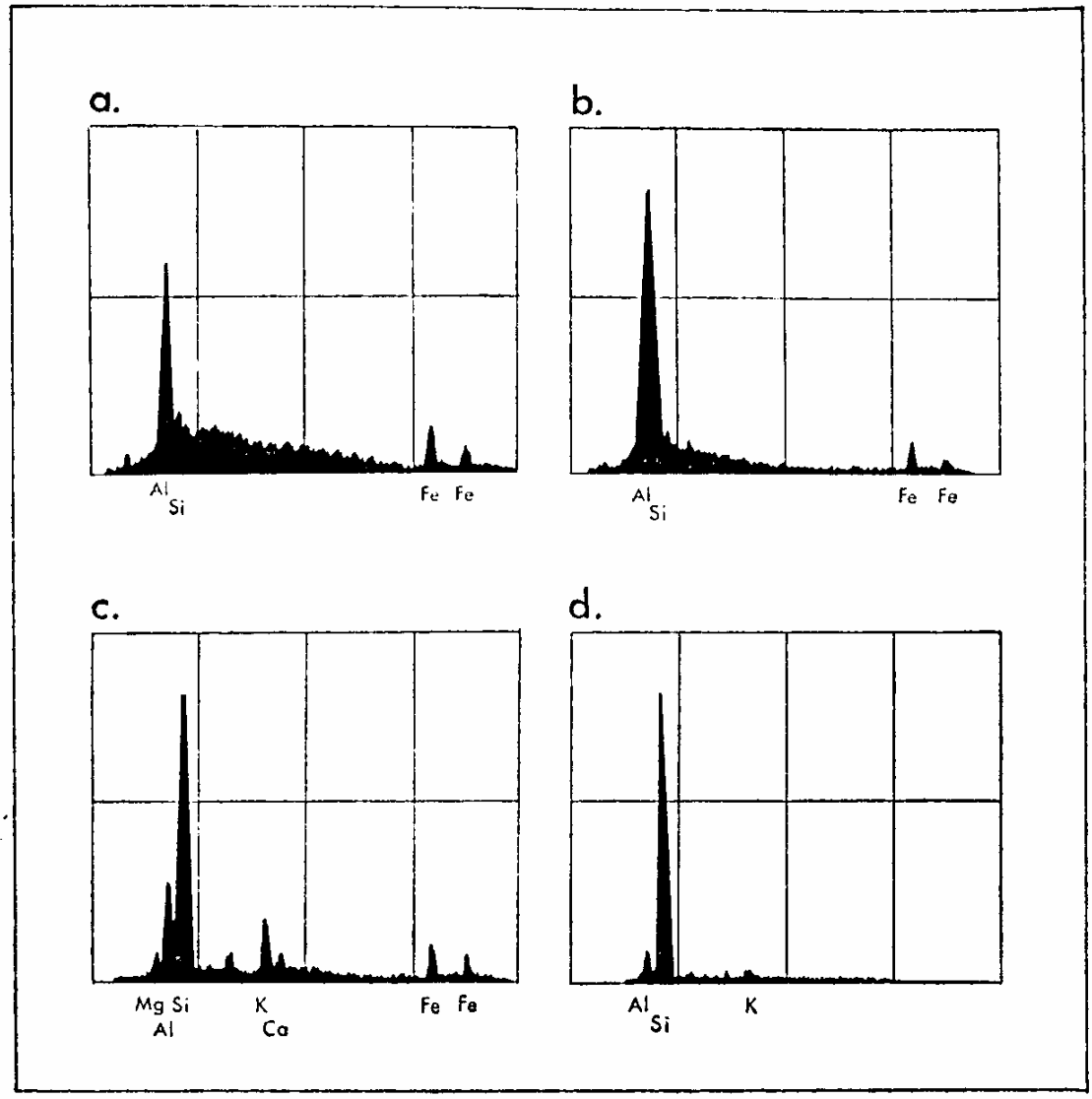

FIGURE 4.

normally Al-rich (Fig. 4a), but analyses of glauconite infillings (pale green in colour) of foraminiferid tests from the same stratigraphic level are Al-poor, and Si-rich (Fig. 4c). These chamber infillings, which must have been deposited during burial or in early diagenesis must represent a spectrum of the elements available in the groundwater at that time. In the same sediment, sponge spicules (pale green in colour) (Fig. 4b) were found to have been replaced by extremely Al-rich, Si-poor, glauconite. Analyses of an assemblage of ostracods (Fig. 4d) extracted from a chert give a profile of almost total silica. These four sets of analyses indicate the extent of diagenetic redistribution which has taken place in the sediment. The originally siliceous spicules are now a silica-poor glauconite, while the glauconite in the foraminiferid tests appears to contain a high percentage of silica.

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The glauconite grains in the sediment appear to be Al-rich, and seem to have taken no part in the redistribution processes. While not wishing to suggest that the cherts are the felted sponge masses of Hinde (1885), it is possible that large scale elemental redistribution (particularly of silica) took place within the sediment during the early phase of diagenesis.

ACKNOWLEDGEMENTS. The author wishes to thank the scanning electron microscopy unit of the University of Newcastle upon Tyne for the geochemical analysis. Prof. J. E. Hemmingway is also thanked for his advice on the problems of chert formation.

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THE ALLER GRAVELS : LOWER TERTIARY BRAIDED RIVER DEPOSITS IN SOUTH DEVON

by R. A. Edwards

Abstract. The Aller Gravel occurs in close association with the Bovey Basin in south Devon. The formation occupies the stratigraphical interval between the Cenomanian Upper Greensand and the Oligocene Bovey Formation, and is probably Eocene in age. At Aller it reaches a maximum thickness of around 20 m and comprises brown and grey flint gravels and sands with subordinate white silty clays and silts, which are interpreted as having been deposited in a braided fluviatile environment. The formation is correlated with the lithologically similar fluviatile division of the high-level flint gravel sequence capping the Haldon Hills east of the Bovey Basin (the Buller's Hill Gravel).

1. Introduction

The Aller Gravel consists of Lower Tertiary flint-bearing gravels and sands occurring in close association with the Bovey Basin in south Devon. The purpose of this paper is to outline the stratigraphy and sedimentary features of the formation, to indicate the likely environment of its deposition, and to discuss the relationship between the Aller Gravel, the Upper Greensand and the high-level flint gravels capping the Haldon Hills east of the Bovey Basin.

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De la Beche (1839), the earliest writer to discuss the stratigraphical position of the Alter Gravel, considered that it rested on the Upper Greensand and was in turn overlain by the "clays and sands of the Bovey Deposit." Woodward (1876) and Reid (1898), however, concluded that there was no Upper Greensand associated with the Bovey Basin, and that sediments of Upper Greensand type were part of the flint gravel sequence. Both writers correlated the low-lying marginal gravels of the Bovey Basin with the Haldon gravels, which Reid referred to the "Bagshot Series." Similarly, Jukes-Browne (1904) correlated the Bovey and Haldon flint gravels, and stated that the two deposits could be "traced from the basin up the slopes towards the Haldon Hills," the "Bovey Beds" having "apparently been let down into a deep syncline by post-Eocene movements."

Clayden (1906) made a series of perceptive observations on Bovey Basin stratigraphy, although his conclusions were hampered by the facts that the Bovey Formation was still considered to be Eocene in age (Gardner 1879), and the existence of Upper Greensand in the Bovey Basin area was thought unlikely. He considered that the (Alter) gravels dipped "as if they formed the floor of the Bovey Deposit," and stated that "Had the Bovey Beds been of Oligocene age, it would be easy to suggest that the whole basin ' was formed after the plateau gravels had been produced . . . . ". He had in mind, therefore, a direct correlation between the Alter and Haldon flint gravels, with subsequent changes in level between the deposits consequent upon the formation of the Bovey Basin.

Reid (in Ussher 1913), rejected the correlation of the Alter Gravel with the Haldon gravels, since he considered that it was a marginal facies of the "lacustrine" Bovey Formation, and in part derived from the older Haldon gravel sheet. He concluded (1913: 116) that "There is no evidence of any former continuity between these low-lying marginal gravels and those capping the (Haldon) plateau about 600 feet higher."

The term 'Alter Gravels' was introduced by Vachell (1963) for the flint gravels typically developed near Alter, Newton Abbot. However, Vachell considered that the gravels "normally overlie the Bovey Beds and in some localities overlap them to lie unconformably on the Permian . . . . the Alter Gravels are redeposited material derived by erosion from the Upper Greensand and Haldon gravel of the Haldon Hills."

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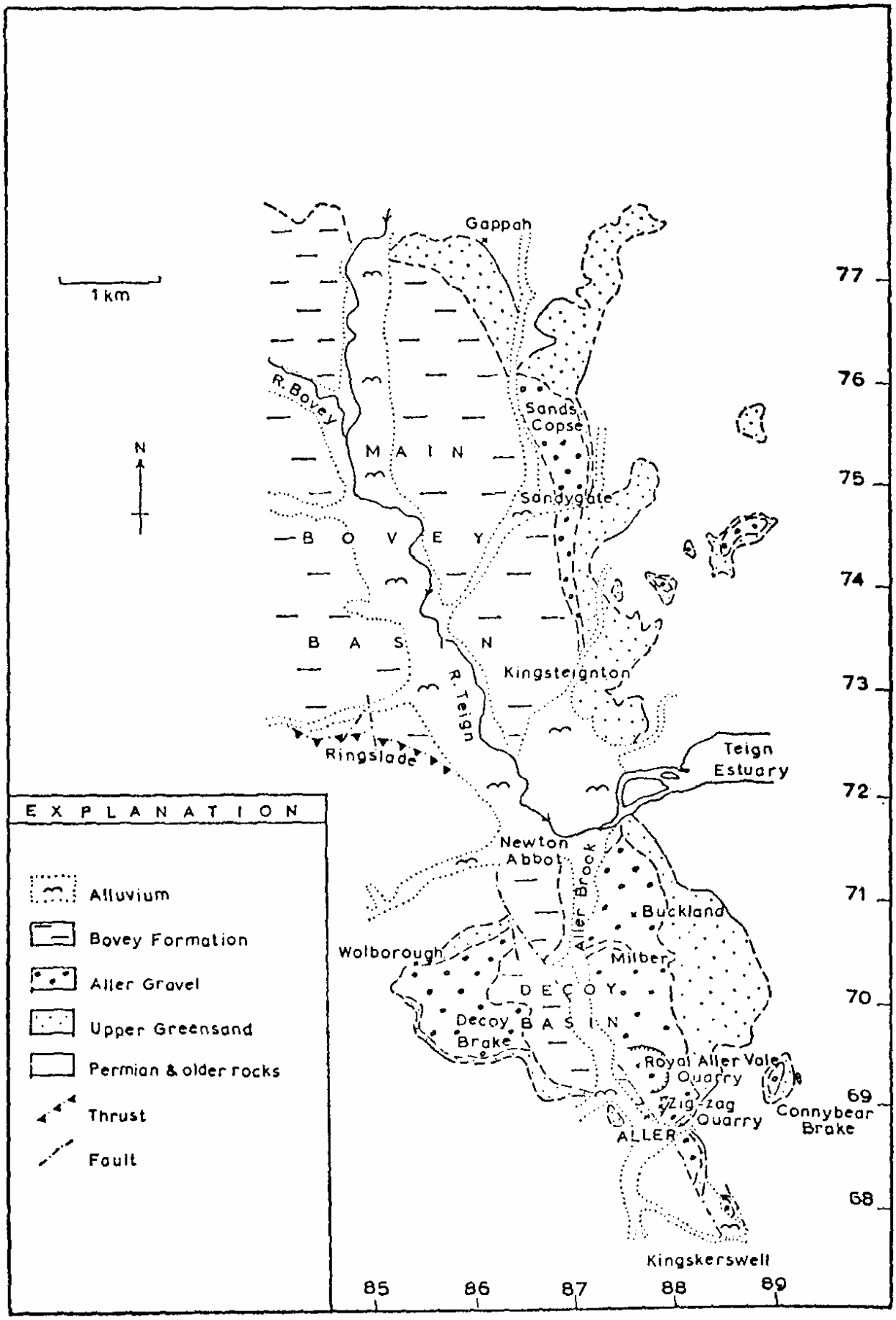

FIGURE 1. Geological map of part of the Bovey Basin

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In a recent study of the geology of the Haldon Hills, Hamblin (1969) rejected a correlation of the Aller and Haldon gravels, and suggested that the Aller Gravel was deposited during or after the formation of the Bovey Basin.

2. **Stratigraphy**

The Aller Gravel, which crops out along the eastern side of the main Bovey Basin and around the Decoy basin south of Newton Abbot (Fig. 1) is a lithostratigraphical unit of formational status. The gravels normally rest directly and unconformably on the Upper Greensand, which is a unit of Cenomanian and ?Albian age, consisting of somewhat glauconitic sands, gravelly sands and sandy gravels, with horizons of tabular cherts (Edwards 1970). The Aller Gravel in turn dips beneath the Oligocene Bovey Formation, a unit of clays, sands and lignites about 1200 m thick, which forms the greater part of the infill of the Bovey Basin.

The pre-Bovey Formation age of the Aller Gravel, previously suspected from general stratigraphical considerations (Edwards 1969), was confirmed by a shallow borehole drilled at Higher Sandygate [SX 8672 7507], near Kingsteignton, by the Institute of Geological Sciences. The borehole passed through pink-mottled clays and sands typical of the lower part of the Bovey Formation, before penetrating flint gravels of Aller type. Hence, the Aller Gravel pre-dates at least part of the Bovey Formation, and since it contains transported Chalk flints and rests on Upper Greensand, is probably Eocene in age.

Where Aller Gravel has been re-worked so that flint gravels rest, for example, on the Bovey Formation or even on true Aller Gravel, it is difficult to distinguish between displaced and *in situ* Aller Gravel. However, Scrivener and Beer (1972) appear to have achieved a distinction within the flint gravel sequence of the Sands Copse area, part of which they have correlated with the 'head' gravels overlying the Bovey Formation, on the basis of lithology, heavy mineral content and distribution.

The most northerly outcrop of Aller Gravel is at Sands Copse (Fig. 1), where flint gravels have been worked in a shallow elongate pit to the east of the Sandygate-Gappah road. The gravel sequence is probably less than 7 m thick, and, as noted above, part may be

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younger than Aller Gravel *sensu stricto.* To the south, at Sandygate, boreholes have shown thicknesses of at least 10 m of Aller Gravel.

Little is known of the extent or thickness of the formation in the Kingsteignton area, owing to the lack of exposure there. North-east of Kingsteignton, three outlying patches of Cretaceous and Tertiary sediments rest on Devonian limestones. The mapping of boundaries in these deposits is made difficult by the effects of solution in the underlying carbonates, but in at least two of the outliers it seems clear that Aller Gravel rests on Upper Greensand. The sub-Cretaceous unconformity in these outliers rises from 300 ft (91 m) O.D. in the most south-westerly, to about 350 ft (107 m) O.D. in the most north-easterly. This contrasts with the relatively steep (7°) west-south-westerly dip of the base of the Upper Greensand in the Kingsteignton area.

The Aller Gravel crops out most extensively and reaches its maximum thickness of some 20 m in the area south of Newton Abbot. On the eastern side of the Aller Brook valley, the outcrop extends from north of Buckland, south through Milber to Aller and Kingskerswell. At Aller, the gravels are worked in two open pits: the Royal Aller Vale Quarry, and the smaller Zig-zag Quarry ; the sections exposed in these pits are the type for the formation. In this eastern area, the gravels dip to the west at between 4° and 10°.

West of the Aller Brook, the Aller Gravel rests on Upper Greensand on the southern slopes of Wolborough Hill, and dips to the SSE ; in the Wolborough area, it is poorly exposed in old pits 100 m WSW of Wolborough church. South of Wolborough, the outcrop of the formation swings round eastwards into the Decoy Brake area, where in old pits in the south of the Brake the gravels dip at 12°-15° NNE. Mapping in the South Quarry area [SX 8689 6958] indicates that the Aller Gravel is missing, having been either over-lapped by the Bovey Formation or faulted out.

The areas east and west of the Aller Brook are probably separated by a NW to NNW trending fault running from Newton Abbot to Aller, and thence to the coast at Torquay. This fault is the southerly continuation of the Sticklepath-Lustleigh Fault.

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No outcrop of Aller Gravel is found along the western margin of the main basin ; if the gravels are present beneath the Bovey Formation in this area their absence in outcrop is explicable by the effects of a prominent western boundary fault. In the northern part of the main basin, the Aller Gravel was possibly removed prior to the deposition of the Bovey Formation, or was overlapped by it, or was never deposited.

Along the eastern part of the southern margin of the main basin, the sediments at the base of the Bovey Formation are concealed by the Ringslade Thrust. Farther west, a unit of gravels (the Staplehill Gravel) crops out, which is lithologically distinct from the Aller Gravel, primarily in the almost complete absence of flint clasts. The stratigraphical relationships of the Staplehill Gravel are not well known ; it may represent a flint-free facies of the Aller Gravel, or, perhaps more likely, a gravelly marginal facies of some part of the Bovey Formation sequence.

3. Sedimentary features and environment of deposition

The Aller Gravel comprises gravels and sands with subordinate silts and silty clays. The coarse sediments are grey or brown in colour, while the silts and clays are generally white or light grey.

The gravels contain a wide variety of phenoclasts, the most important of which are Chalk flints. These show a moderate degree of abrasion, which has affected especially the prominent horns. The surfaces of flint cobbles may show small semi-circular fractures ('chatter markings'), which are presumed to have been formed by the mutual impact of cobbles during transport. Other components of the gravels include abraded blocks of Upper Greensand chert, vein quartz, tourmaline and quartz-tourmaline rock, light grey Lower Carboniferous chert, grey Upper Carboniferous (Crackington Formation) sandstone, ?dolerite, metadolerite, and hornfels and other dark grey fine-grained rocks which are considered to have originated from the aureole of the Dartmoor Granite. These components point to derivation from a northerly or north-westerly source.

Sedimentation units of the Aller Gravel are generally lenticular in cross-section, and bounded by curved to straight erosional surfaces. Marked lateral and vertical changes in grain size are common. In places, piles of channel forms may be seen resting

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on each other with erosional junctions. The channel sediments are generally poorly sorted and may show large scale cross-bedding. Fine-grained deposits are restricted in extent and usually take the form of small lenticular silt or clay bodies. Clays also occur as sub-round intraformational clasts set in a gravel or sand matrix.

The sedimentary features noted in the Aller Gravel correspond most nearly to those described from modern braided fluviatile sediments (e.g. Doeglas 1962 ; Williams and Rust 1969), which are characterised by numerous channels of coarse sand and gravel, with rare fine-grained sediment representing deposition in depressions or channels on the floodplain. The absence from the Aller Gravel of fining-upward sequences with point-bar sands and extensive overbank fine-grained deposits contrasts with the meandering stream model. Thus the gravels were probably deposited on an extensive floodplain extending for a considerable distance eastward and southward from the Dartmoor Granite upland.

4. Correlation of the Aller Gravel and Haldon gravels

As noted above, a number of authors have considered that the flint gravels of the Bovey Basin correlate directly with the flint gravels capping the Haldon Hills. More recently, others (Vachell 1963 ; Hamblin 1969) have followed Reid (in Ussher 1913) in rejecting such a correlation. The present author supports a direct correlation of the Aller Gravel and the fluviatile facies of the Haldon gravel sequence. It is considered that the present low-lying position of the Aller Gravel/Upper Greensand sequence is a consequence of the tectonic subsidence associated with the formation of the Bovey Basin, and that it is this tectonically explicable difference in levels between the flint gravels and Upper Greensand exposed in the Bovey Basin area and on the Haldon Hills which has in part obscured the otherwise clear evidence of their correlation.

Hamblin (1969) recognised three divisions within the flint gravel sequence of the Haldon Hills:

(i) The "Residual Gravel" (Tower Wood Gravel), a residual flint gravel of unabraded flints in a clay matrix, formed by the *in situ* solution of Cretaceous Chalk.

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(ii) The "Haldon Fluvial Gravel" (Buller's Hill Gravel), a set of gravels with abraded flint the dominant component, also containing "vein quartz, tourmaline rock, quartz-tourmaline, quartzite, and baked aureole rock in a matrix of sand and clay." This gravel contains clay bodies, the "Haldon Clay Deposit" of Hamblin, which he considered to have "attained their present distribution, as bodies within the Fluvial Gravel, as a result of cryoturbation and solifluction during the Pleistocene" (Hamblin 1969: 311). This conclusion was based on differences in clay mineralogy between the clays of the Haldon Clay Deposit and the clays occurring interstitially in the gravels.

(iii) The "Haldon Solifluction Gravel" (Head Gravel), which was believed to have been formed by the solifluction of the older flint gravels, so that it now mantles the slopes of the Haldon Hills.

The Tower Wood Gravel and the Haldon Solifluction Gravel are not represented in the Bovey Basin, although there may have been some Pleistocene re-working of the Aller Gravel. We are concerned here with the evidence for correlating the fluviatile division of the Haldon gravels with the Aller Gravel, which is presented below.

(a) Stratigraphy : The Buller's Hill Gravel rests on Cretaceous sands of Upper Greensand facies, as does the Aller Gravel. These two sequences are only 2 km apart at their closest point between Great Haldon and Ugbrooke, although there is a significant difference in altitude between the two areas. The similarity and proximity of the Haldon and Bovey sequences argue for their correlation.

(b) Lithology: There are marked lithological similarities between the Aller Gravel and the Buller's Hill Gravel, both units comprising fluviatile flint gravels with similar components and similar sedimentary structures.

(c) Position: The Aller Gravel and Upper Greensand occur along the eastern margin of the Bovey Basin and between Bovey and Haldon in positions such as to indicate that they lie on a westward dipping surface connecting the two areas. It is probable that the outliers such as that at Connybear Brake are remnants of a once continuous sheet which formerly connected the Upper Greensand and flint gravels of the Bovey Basin with those of the

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Haldon Hills. The tectonic subsidence associated with the formation of the Bovey Basin has deformed this sheet so as to give it a westerly dip ; subsequent erosion has destroyed the former continuity between the now low-lying Upper Greensand/flint gravel sequence and that of Haldon, which is left at a higher level. Had erosion not destroyed the connecting outcrops, it would be possible to trace a single Upper Greensand/flint gravel sheet between Haldon and Bovey, and the idea that the gravels of this sheet did not correlate would not be entertained.

As a result of the arguments presented above, it is considered that the fluviatile flint-bearing gravels of the Bovey Basin and Haldon Hills are a product of the same Lower Tertiary sedimentation episode.

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STRUCTURE OF THE DEVONIAN LIMESTONE AT BRIXHAM

by D. K. Smythe

Abstract. The Devonian limestones at Brixham show a structure which is transitional in style between the overturned folding of the South Hams region and the thrusting found east of Dartmoor, and can be accounted for by a single phase of deformation.

1. Introduction

The Middle to Upper Devonian limestones at Brixham are well exposed, and occur in a structurally simpler setting than those to the north. This paper is based upon structural mapping of the area south of Tor Bay and east of the River Dart.

2. Stratigraphy and structure

The E-W trending upright anticline of St. Mary's Bay duplicates the outcrop of the limestones, and 150m or so- of grey Eifelian shales are exposed in the core of this double-hinged fold (Fig. 1). As shown in Figure 2, the northern flank of the anticline exposes 300m of limestone, divisible into three units of roughly equal thickness. The uppermost unit, of poorly-bedded limestone with little crinoid content, passes laterally into the massive algal stromatoporoid reef limestone of Berry Head. The middle unit, of thin-bedded crinoidal limestone, reappears from below the massive Berry Head limestone in the core of the upright, angular, ENE trending Shoalstone Beach anticline. The lowermost unit comprises bands of crinoidal limestone 10-20cm thick, separated from each other by 1-2cm thickness of red shales. Slaty cleavage in the shales and fracture cleavage in the limestones (except where they are massive) trend generally E or ENE, and dip S at 30°-50°.

Although the bulk of the limestone ranges from late Eifelian to Frasnian in age, the absence of Eifelian shales south of the Sharkham Point outcrop presents a problem which Ussher (1903) overcame by postulating an E-W thrust to separate the limestone outcrop from that of the Staddon Beds. The evidence of critical localities is discussed below.

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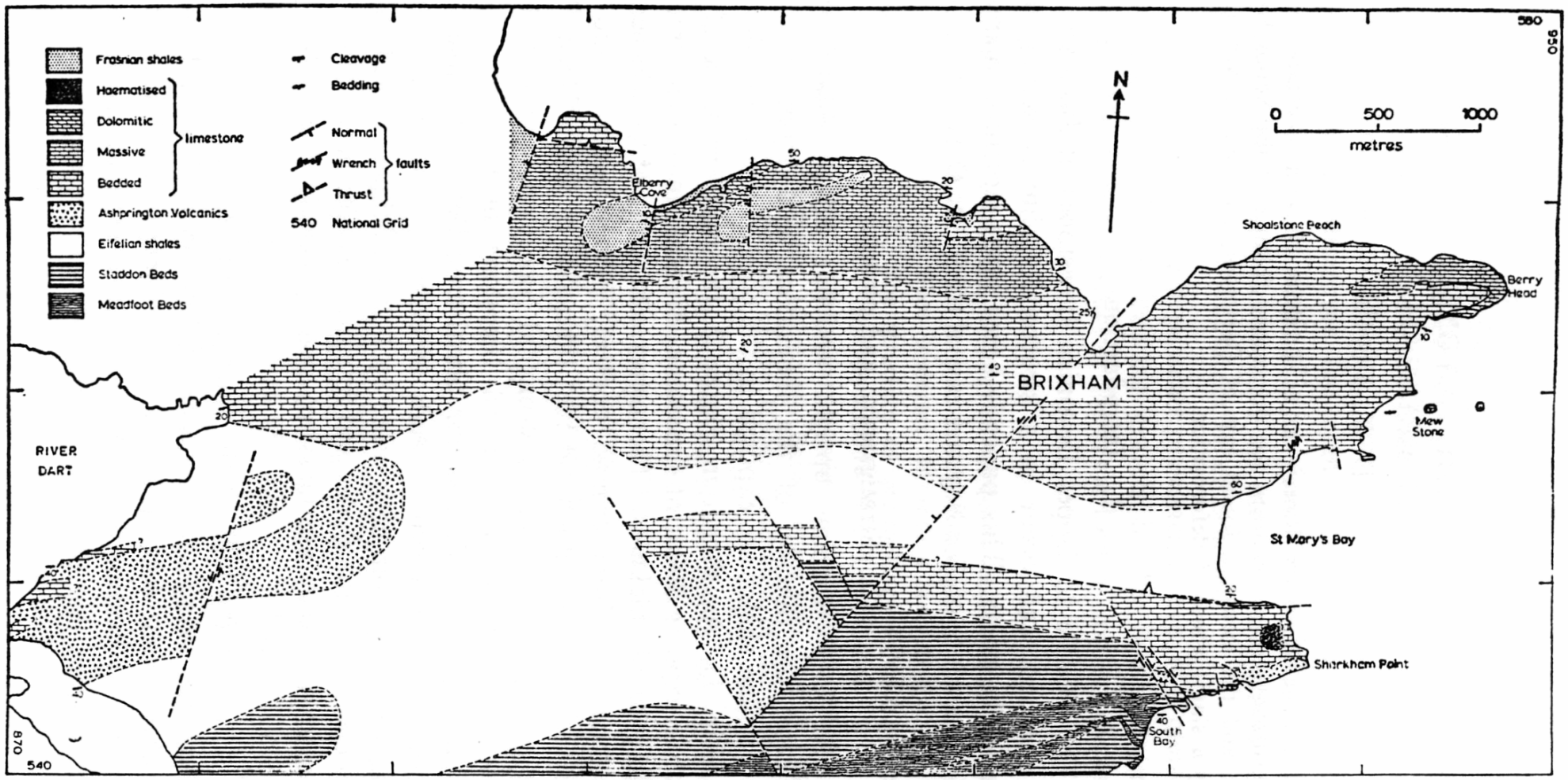

FIGURE 1. Geology of the Brixham District

(*a*) *South Bay* (SX 929543)

In the N-S cliff section uppermost Meadfoot Beds (steely-grey and black slates) and lowermost Staddon Beds (lilac-red sandstones and slates) are affected by tight folds with horizontal E-W axes and with axial planes which dip S, parallel to the slaty cleavage, at about 40°. To the east of the Cove (SX 93005438) the Staddon Beds are cut off by a large NW-SE trending oblique-slip dextral wrench fault dipping SW at about 80°. On the north-east side of the fault Staddon Beds forming a small islet (SX 93045438) can be traced northwards through 15-20m of transitional red shales, with thin limestones, into thickly-bedded crinoidal limestone dipping south at 50-60°. The limestone here (SX 93025443) has yielded a fauna of icriodid conodonts characteristic of the lower Eifelian (S. C. Matthews, pers. comm.).

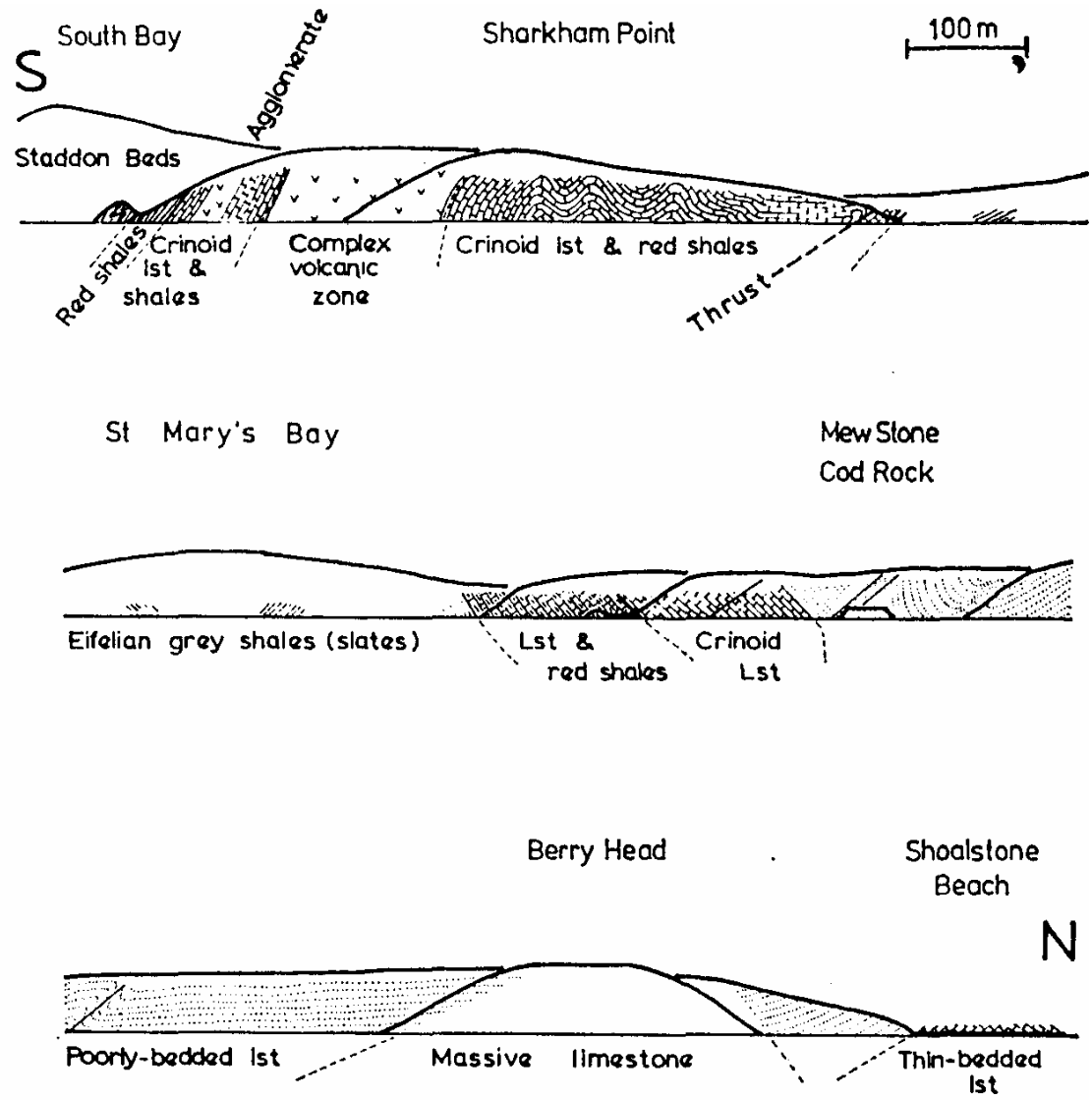

FIGURE 2. Section along the eastern coastline from South Bay to Shoalstone Beach. No vertical exaggeration.

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(b) Coast west of Sharkham Point (SX 937546)

A highly altered sequence of tuffs, agglomerates and shales, the most easterly outcrop of the Ashprington Volcanic Series, overlies the southern outcrop of the limestone, but structurally underlies the limestone at South Bay, which contains two agglomerate beds. The structure is complex, due principally to late normal faulting. A certain degree of symmetry in the distribution of beds about an E-W trending surface suggests, however, that the volcanics occupy a synclinal fold closure, flanked on either side by limestone. The fold axial surface dips south at 50-60° near South Bay, but is vertical and cut by several faults at Sharkham Point.

(c) Coast north of Sharkham Point (SX 936548)

Approximately 50m of crinoidal limestones and shales, locally rich in rugose corals, and in places almost completely replaced by disseminated haematite, are folded into fairly open upright E-W folds. Along the south side of St. Mary's Bay the roughly horizontal limestone is overturned above a large E-W trending thrust plane which dips south at 30-40° (SX 93565488). The beds below the thrust (transitional thin shaly limestones overlying the Eifelian shales) are, however, parallel to the thrust plane.

(d) Brixham town (SX 925560)

A large NE-SW trending fault is postulated to run through Brixham harbour (Fig. 1). It separates the mainly horizontal limestone tract to the east from the outcrop of flat-lying recumbently folded limestone to the west. In the vicinity of this fault, both bedding and cleavage in the limestones and shales swing parallel to its trend.

(e) Elberry Cove area (SX 903570)

The massive limestone along the south coast of Tor Bay, west of Brixham, shows recumbent minor folds trending E-W and facing north. From Elberry Cove (SX 904570) to Ivy Cove (SX 908571) the limestone (here dolomitised) structurally overlies bright red shales. Goniatites and conodonts from the shales and limestones indicate that they are of Frasnian age (House 1963, Tucker and van Straaten 1970). The red shales overlying the limestone, 50m above, on Churston golf course (SX 908570) may be the same age.

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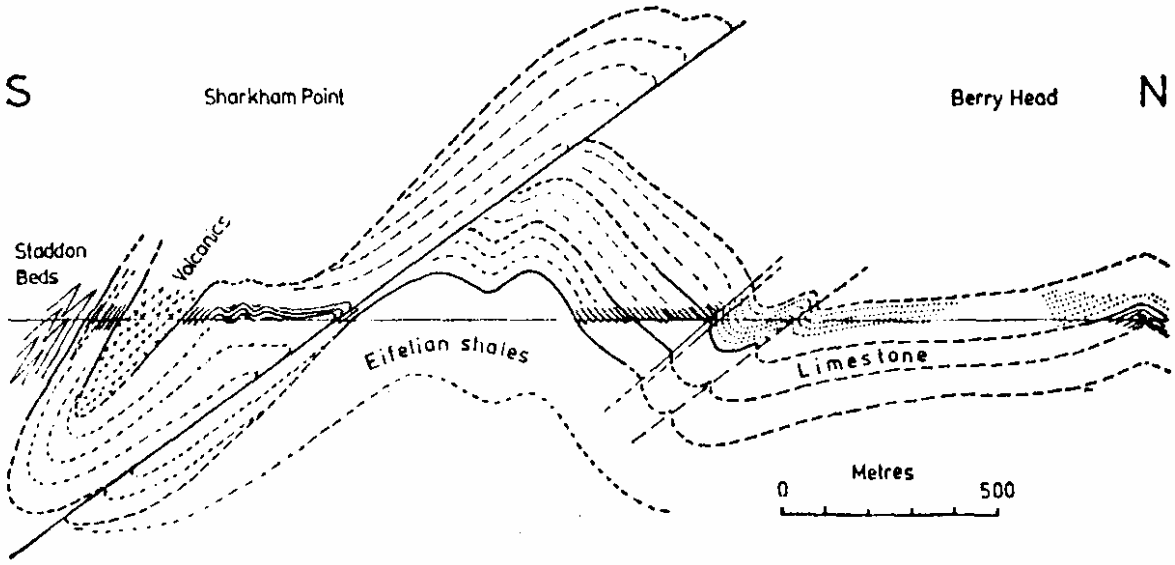

FIGURE 3. Structural interpretation of the section from South Bay to Shoalstone Beach. No vertical exaggeration.

3. Structural interpretation

The simplest interpretation of the geology of the Sharkham Point-South Bay area is shown in Figure 3. For the purposes of geometrical reconstruction, similar folding has been assumed, as it is difficult to assess the relative importance of similar and concentric folding during the deformation. However, any errors in geometry caused by this assumption are likely to be less than those involved in reconstruction of the stratigraphic sequence. As the Sharkham Point thrust cuts across. a large fold, with the dip of the rocks above and below differing by 40-50°, a large displacement is inferred. This is confirmed by the scale of the 'drag' above the thrust plane, which bends the beds with a radius of 5-10m through an arc of 150°. Furthermore, the apparently different Eifelian successions at St. Mary's Bay and South Bay can be more readily reconciled if thrusting as well as folding has brought them close together.

It is tempting to imagine that the Brixham harbour fault was initiated as a sinistral wrench, complementary to the major dextral wrench faults of the Dartmouth-Kingswear district, but there is no evidence for any hade, and the downthrow is opposite in sense to that of the minor sinistral wrench faults. As the axial plane of the St. Mary's Bay anticline here dips south at about 70°, a downthrow to the north-west and a sinistral horizontal component of displacement, each of about 100m, is estimated.

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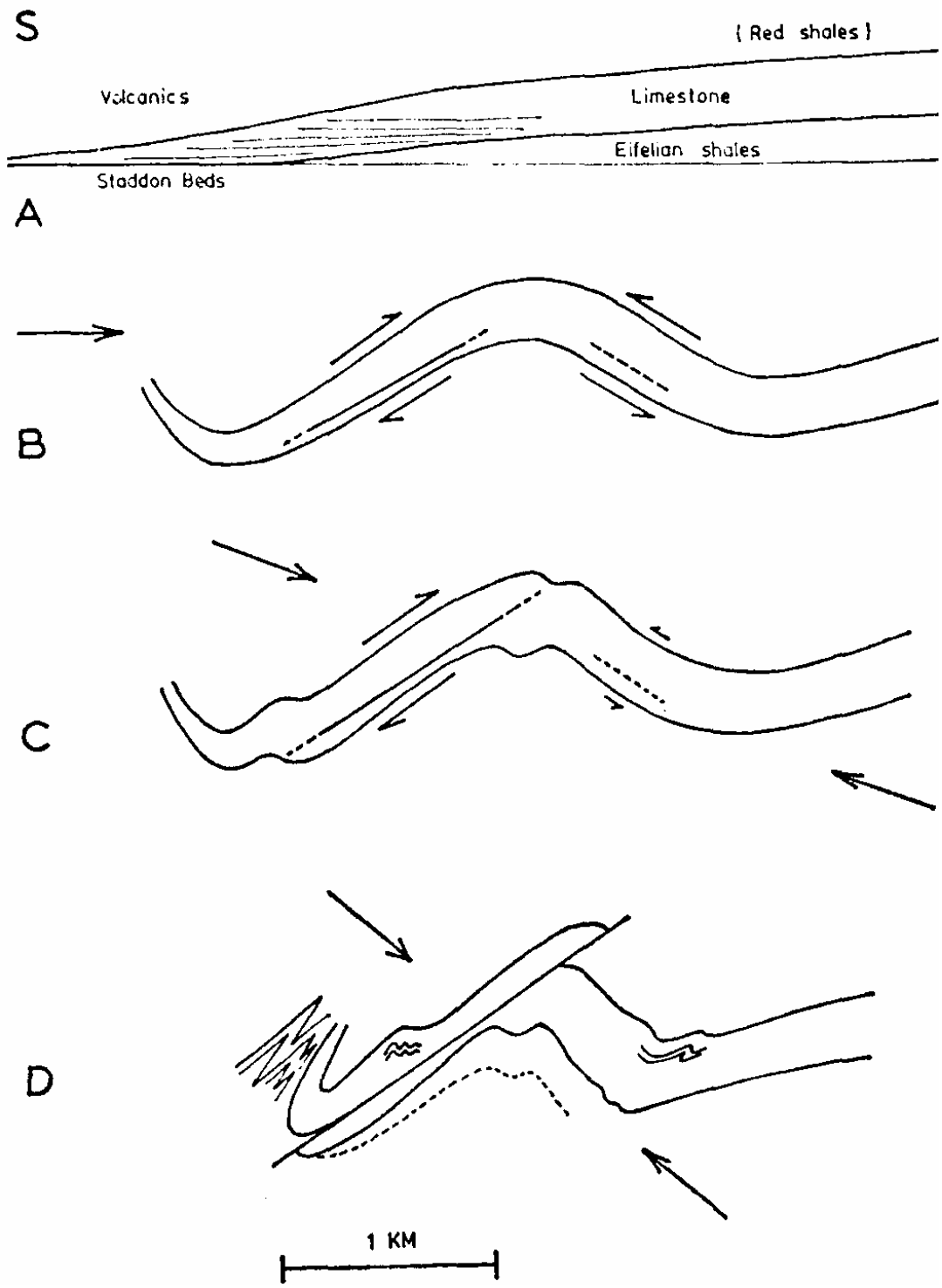

FIGURE 4. Development of the major folds and Sharkham Point thrust. Full-headed arrows show components of maximum principal compressive stress, half-headed arrows show components of shear stress.

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Towards the River Dart the St. Mary's Bay anticline becomes progressively tighter and more overturned to the north, more similar in style to the large folds in the Lower Devonian to the south. On its southern flank the limestone has been almost entirely replaced by several hundred metres of volcanics, which probably underlie stratigraphically the interfolded thin shaly limestones.

4. Geological history

The Brixham limestone is strongly diachronous ; in early Eifelian times it is laterally equivalent to the grey shales further north, and later on is coeval with the volcanics to the west and south (Fig. 4A). The Sharkham Point area may thus have been the original site of a swell separating shallow water to the north from deep water to the south. The apparent migration northwards of the swell and growth of a reef in the Berry Head area is consistent with differential subsidence of the Middle Devonian continental shelf, the greater relative downwarping being in the south, but with deposition keeping pace with the subsidence (Dineley 1961).

Rheologically the limestone will behave as a competent wedge set in incompetent shales and volcanics. The maximum principal compressive stress during the main phase of folding, which probably occurred in the Lower Carboniferous (Sanderson and Dearman 1973) must have been initially horizontal, and secondary shear stresses would have been set up on the flanks of the large open flexural folds first formed (Fig. 4B). As the maximum stress rotates during regional tectonic transport (Smythe 1971) and becomes progressively inclined to the north, the secondary shear stress on the northern flank of the main anticline is reduced, whereas that on the southern flank is increased (Fig. 4C). Bedding surfaces provide a ready plane of weakness for thrusting, while the displacements toward either end of the thrust are taken up by folding. The culmination of the main phase of deformation is the development of medium-scale and minor asymmetric folds over-turned to the north, and the growth of the slaty and fracture cleavages parallel to the axial planes of these folds (Fig. 4D).

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The oblique-slip wrench faults were possibly formed at about this time, as the orthogonal stress system for the faulting is identical to that for the folding, except that σ_2 and σ_3 are interchanged. This interchange results naturally from considering the rock either as an elastic solid or as a viscous fluid ; deformation then occurs by faulting or by folding respectively. As all the wrench faults of the area, including the large dextral wrench faults east of Kingswear, are oblique-slip and explicable by the stress system shown in Fig. 5, it seems possible that the major dextral wrenches of the area, throughout perhaps the whole of SE Devon, originated during the Variscan orogeny, and not in the Tertiary, as suggested by Dearman (1963), although the major displacements along most of them probably did occur in the mid-Tertiary.

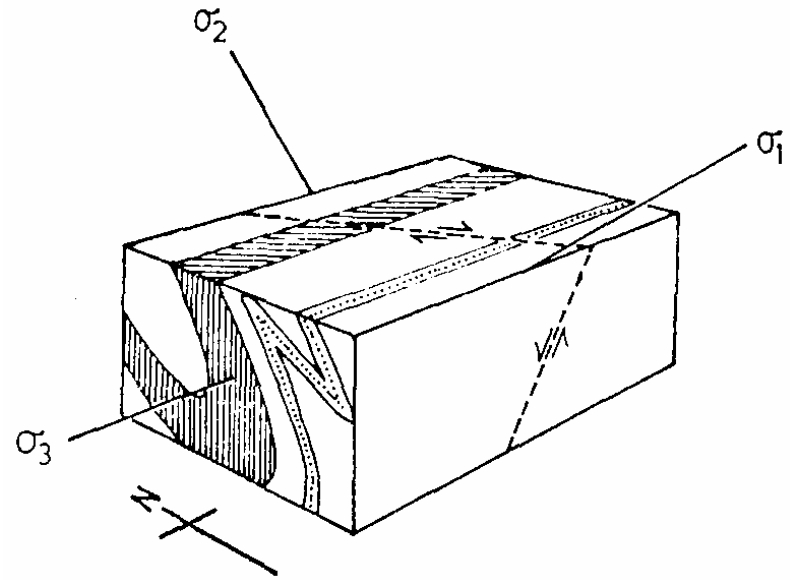

FIGURE 5. Block diagram of the stress system resulting in oblique-slip dextral wrench-faulting at South Bay, looking down to the NE. Folding in limestone (vertical ruling) and Staddon Beds (dotted) shown diagrammatically. σ_2 is parallel to the fault plane.

The structure of the district illustrates that the change from overturned folding in south Devon to recumbent folding with large-scale thrusting further north (Simpson 1969, Sanderson and Dearman 1973) is transitional rather than abrupt, and that both styles of deformation can be accounted for by a single stress field, which was more intense in the north.

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