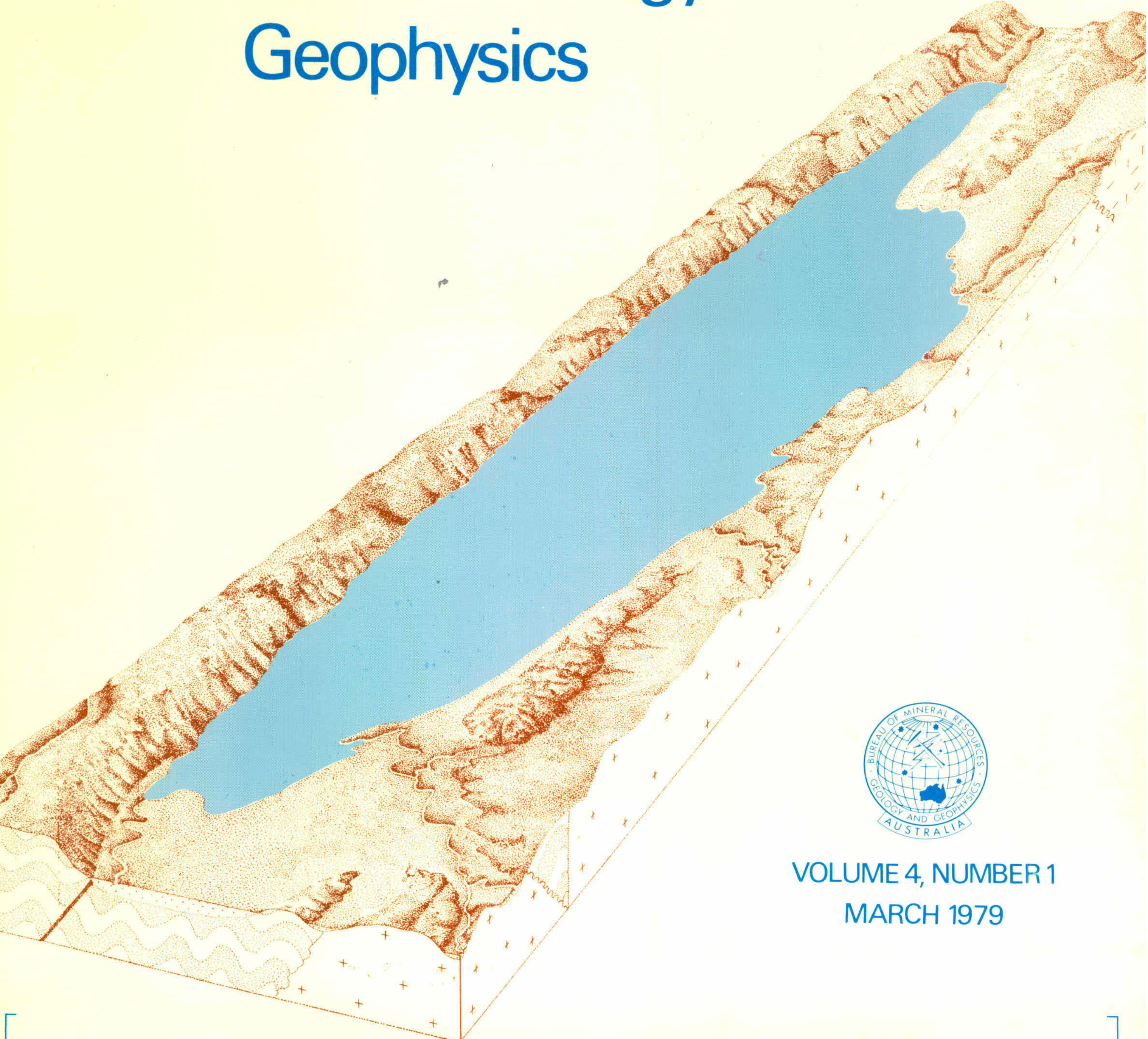
BMR JOURNAL of Australian Geology & Geophysics



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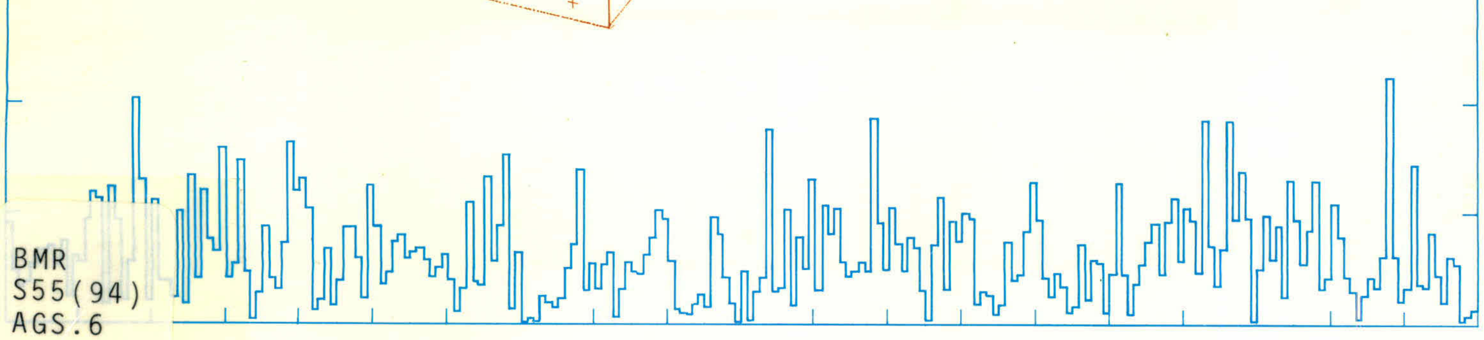

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Front cover:

A block diagram of Lake George, a closed lake in New South Wales. Marked fluctuations of the lake's water level have been the subject of much speculation. A paper in this issue discusses the water balance of the lake, based on monitoring over a 20-year period. The histogram shows rainfall variation in the same period.

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Regional variations in Australian heat flow

J. P. Cull & D. Denham

Twenty new heat-flow values from widely separated sites on the Australian continent are presented. These are combined with published land and marine data in the area 90-170°E, 0.45°S to produce contour maps of heat flux in the region. For most of the area the heat flux is uniform, and low (30-60 mW m^{-2}), presumably because both the oceanic and continental crusts are comparatively old. However, in eastern Australia the heat flow is significantly higher ($\sim$ 70-100 mW m^{-2}). The new data reinforce this pattern with values from the Canning Basin and Pilbara Block falling in the range 40-50 mW m^{-2}, and those in Queensland and New South Wales in the range 70-100 mW m^{-2}.

The data distribution is still not good enough to delineate accurately the boundaries of the high and low heat-flow regimes, and hence define the boundary of possible Precambrian craton(s) where no outcrop is present.

There is a strong positive correlation between heat flow and P-wave travel-time residuals, and a negative correlation between heat flow and upper-mantle velocity. There is no obvious relationship between heat flow and crustal thickness.

Introduction

Most heat within the Earth is generated through the decay of radioactive isotopes contained in crustal rocks (Roy & others, 1968); of lesser importance are contributions from deeper sources associated with global accretion, core segregation, earth tides, and core/mantle rotation (MacDonald, 1963). This heat is generally conducted to the surface of the Earth at rates of about 60 mW m^{-2}; this figure implies a total heat loss exceeding 10^7 MW. No other process within the Earth can supply similar quantities of energy; consequently models of global tectonism are constrained by observations of surface heat flow.

Oceanic heat-flow patterns are consistent with models of mantle convection (Turcotte & Oxburgh, 1969). However, there is no well established model for continents. Regions of anomalous heat flow may be generated by convection cells (and possibly plumes) acting at the base of the lithosphere. Resulting thermal stresses may then be responsible for major structural features including fold mountains, rift valleys, and active volcanism. These features can be delineated from seismic data using velocity/density systematics. However, seismic velocities depend not only on the composition of rocks but also on their temperatures. Anomalous seismic data may be caused by changes in rock composition, structure, or temperature; the effect of each must be resolved before the convection hypothesis can be assessed in detail.

On a local scale, geothermal surveys have been used to locate fault systems, dome structures, coalfield boundaries, and mineral deposits undergoing exothermic reactions due to hydration or oxidation. In addition the thermal history of sedimentary basins must be considered in assessing petroleum prospects, since the level of maturation of hydrocarbons is a function primarily of time and temperature.

Geothermal energy can also be considered as a national resource. At plate boundaries geothermal activity is intense, and hydrothermal systems of high enthalpy are common. Steam and hot water may be readily available for use in power turbines. Elsewhere, geothermal activity may not be obvious. However, low enthalpy systems are found in many sedimentary basins, often associated with volcanics containing residual heat. Hot water can be extracted in these regions for domestic and industrial use. Heat-flow data are required to assess energy content and extraction rates for resources of this type.

Maps are presented in this paper to indicate the number of heat-flow data and the major trends of the geothermal field in Australia. Possible correlations are suggested relating heat flow to other types of geophysical data used in characterising the crust and upper mantle.

Sources of data

Surface heat flow Q (W m^{-2}) is calculated from the expression

$$Q = \beta\lambda \qquad \text{(1)}$$

where β (mK m^{-1}) is the geothermal gradient, and λ (W m^{-1} K^{-1}) is the thermal conductivity of the rocks in which the gradient is established.

Near-surface temperatures on land are determined primarily by climatic conditions including seasonal variations in surface temperature. Consequently geothermal gradients are usually measured at depths greater than 100 m. Data have been obtained in tunnels and mines, but boreholes are required for most areas. Unfortunately, the normal thermal regime is disturbed by the circulation of fluids during drilling. Equilibrium may not be regained for periods up to ten times the duration of drilling (Bullard, 1947); consequently casing may be required to prevent collapse prior to logging for absolute values. Thermal gradients can then be verified using multiple runs to detect water movement and instrument error.

If casing is not available differential temperatures can be measured with little error shortly after drilling is completed (Jaeger, 1961); furthermore, corrections can be made to bottom-hole temperatures measured during pauses in drilling (Oxburgh & others, 1972). However, it is not possible to verify data obtained with these transient techniques. Furthermore, logging facilities must be maintained at each drilling site to ensure that rig time is kept to a minimum. Costs of drilling and casing, together with the logistics of logging are therefore major factors in limiting the number of heat-flow data available from sites on land.

In deep oceans the temperature of the sea floor is almost constant. There are no climatic pertubations, and

Location	Lat. (°S)	Long. (°E)	elev. (m)	D (m)	β (°C/km)	λ (Wm-1K-1)	N	Q (mW m-2)
Noonkanbah	18.11	124.81	120	40-140	31.7 ± 0.7	2.12 ± 0.40	13*	67.2 ± 3.5
Billiluna No. 3	19.51	127.64	450	40-70	14.2 ± 0.8	3.22 ± 0.18	4*	45.8 ± 2.3
Bannerman No. 1	19.95	127.23	450	35-65	39.9 ± 1.6	1.47 ± 0.13	3*	58.7 ± 5.7
Tanami	19.98	129.71	450	40-80	18.6 ± 0.5	1.85 ± 0.33	4*	34.4 ± 6.2
Halls Creek	18.24	127.69	510	35-95	9.4 ± 0.6	2.72 ± 0.01	2*	25.6 ± 1.6
Yuendumu	22.33	131.75	730	0-220	B(1)	B	6	55.8 ± 3.1
St George Range	18.69	125.14	210	0-4400	15.9 ± 1.4	3.94 ± 0.67	11	62.7 ± 4.0
					B	B	11	64.9 ± 4.0
					adopted value			63.8 ± 4.1
Wittenoom 47A	22.34	118.24	720	40-115	7.7 ± 0.1	3.40 ± 0.16	2	26.1
				115-185	9.6 ± 0.2	4.15 ± 2.20	2	39.7
				185-235	22.1 ± 0.3	1.78 ± 0.22	5	39.3
				235-300	11.6 ± 1.9	3.66 ± 1.19	6	42.5
					adopted value			40.6 ± 2.9
Millstream No. 9	21.65	117.02	320	60-100	7.3 ± 0.1	5.03 ± 0.37	4	36.7
				100-145	7.8 ± 0.1	5.10 ± 0.23	4	39.8
				145-175	8.0 ± 0.0	4.72 ± 0.76	3	37.8
				175-230	10.6 ± 0.9	4.17 ± 0.08	4	44.2
					adopted value			40.5 ± 1.3
Tom Price G623	22.75	117.77	950	40-190	B	B	4	45.0 ± 1.9
Mount Morgan 63/3	23.63	150.37	350	60-140	15.60(2)	3.66	5	56.8 ± 1.2
Mount Chalmers MW4	23.29	150.65	100	150-230	23.17	2.69	6	62.3 ± 0.9
Peak Downs Utah 22257	22.28	148.22	200	95-280	35.0	2.34	7	79.2 ± 5.6
					B	B	11	80.7 ± 1.9
				120-200	38.6 ± 0.6	1.98 ± 0.26	6	76.3 ± 4.1
Balfes Creek	20.33	145.85	300	68-230	18.11	4.26	10	75.7 ± 4.0
				61-174	21.22	3.65	9	77.4 ± 4.7
Dugald River	20.31	140.20	300	127-283	20.95	3.50	8	72.0 ± 2.9
Mount Dore	21.78	140.55	360	65-185	32.67	2.90	4	98.2 ± 4.1
Canberra BMR 145	35.30	149.15	605	130-260	25.38	2.89	13	72.5 ± 2.7
Woodlawn Jododex CE12	34.98	149.52	833	78-262	19.33	2.99	12	59.6 ± 1.5
Ardlethan	34.34	146.85	220	100-200	22.55	4.70	5*	105.4 ± 4.4
				60-200	21.16	4.39	7	92.9 ± 5.9
Mount Gambier BMR PEN 31	37.75	140.89	70	71-243	33.57	3.47	19	91.8 ± 3.0

(1) B indicates Bullard reduction of data. (2) Q calculated as function of depth. * indicates that cell method has been used to determine λ. N is the number of samples used for conductivity determinations. Errors quoted in β, λ refer to standard deviation of samples, error in Q refers to standard deviation of mean.

Table 1. New heat-flow data.

geothermal gradients are readily measured with probe techniques (Bullard, 1954). More data are available than on land, but even so, limits are imposed by the cost of ship charter.

Prior to 1950 there were only a few scattered determinations of heat flow—in South Africa, Britain, and North America. However, in 1975 the number of determinations had increased to more than 5000, 30 percent of which were on land. Compilations have been published by Lee & Uyeda (1965), Simmons & Horai (1968), and Jessop & others (1975).

In Australia the first heat-flow data were published by Newstead & Beck in 1953. Since then several workers at the Australian National University (ANU) have published data; by 1977 heat-flow values had been determined for 90 widely distributed sites. Compilations were published by Sass & others (1976), and Lilley & others (1978). Marine data have not been obtained on a routine basis for the Australian region, and values used in this paper were extracted from intermittent surveys conducted by visiting parties from foreign institutes. Data are reported by Sclater & others (1972), MacDonald & others (1973), Erikson & others (1975), Jongsma (1976), and Anderson & others (1977). Most of these results are listed by Jessop & others (1975).

Measurements of surface heat flow are now included in onshore programs conducted by the Australian Bureau of Mineral Resources (BMR). Data have been obtained mainly on an opportunity basis in holes drilled by private companies for mineral exploration, but BMR stratigraphic holes have also been surveyed, and special-purpose heat-flow holes have been drilled to extend the regional coverage. Results for 20 new locations are presented in the appendix. A summary of the data is given in Table 1.

Errors in data reduction

Geothermal gradients are usually calculated from absolute measurements of temperature at depth intervals not less than 5 m. Thermistor probes are commonly used for this purpose, giving a resolution better than 0.001°C (Cull & Sparksman, 1977). The observational errors in computed temperature gradients should therefore be less than 1 percent (assuming common geothermal gradients of about 30 mK m^{-1}). However, there are frequently departures from a linear temperature gradient caused by changes in thermal conductivity, which depend on rock type.

Although conductivities can be measured on a divided-bar apparatus with an accuracy of better than

2 percent (Cull & Sparksman, 1977), usually only a few core samples are obtained from each borehole—consequently sampling errors are introduced in calculating heat flow. In practice, therefore, heat flow is determined for several intervals with linear gradient and the results are averaged to give estimates of statistical error. In general the instrumental errors can be neglected.

If a normal distribution is assumed to describe variations in thermal conductivity in each borehole, a standard deviation can be assigned to the mean of the individual determinations of heat flow. However, these results constitute only a sample of the total population, and the precision is given by the standard deviation of the mean (σ_m), which is related to the standard deviation of the sample (σ) through the expression

$$\sigma_m = \sigma / \sqrt{N} \qquad \text{(2)}$$

where N is the number of determinations constituting the sample. In most cases the variation in thermal conductivity values is the controlling factor in the calculation of σ_m.

Bullard plots

If borehole stratigraphy is well known, heat flow can be determined using estimates of interval resistivity. Assuming that heat flow is constant through all intervals, the temperature at any depth can be expressed as (Bullard, 1939)

$$T_z = T_o + Q \, \Sigma_i (D_i / \lambda_i) \qquad \text{(3)}$$

where D and λ are respectively the thickness and conductivity of each section of uniform lithology, and the depth $z = \Sigma_i D_i$. Conductivities can be assigned to each rock type and a value of Q can be obtained by plotting $\Sigma_i D_i / \lambda_i$ against values of T_z. Conductivities are again subject to sampling errors; these are reflected in the correlation coefficients. Precision can therefore be estimated with regression techniques, giving results equivalent to those discussed above (σ_m in equation 2).

Data compilation

The new data presented in the appendix (see also Table 1), were combined with previously published values (see Introduction for references) to produce the contour maps shown in Figures 1 and 2. At locations such as Mount Isa, Tennant Creek, and Kalgoorlie, where multiple determinations have been made, a best value (usually a simple average) has been adopted for a mean location. Data are available for 103 locations on land, while at sea there are 241 determinations in the region 90-170°E, 0-45°S. However, there are still large unsampled areas, notably on land in the desert regions of Western Australia, in Cape York, and in southern Queensland.

The quality of heat-flow values comprising the data bank varies considerably. Temperatures have been measured in tunnels, mines, flowing bores, and cased and uncased holes of variable depth. Different observations are therefore subject to different types of correction, but, in general, thermal gradients are not much altered. Most errors in heat-flow data result from inadequate determinations of thermal conductivity. Measurements are made with a divided-bar apparatus calibrated with quartz standards to provide consistent results. However, errors are introduced: where the lithology is extremely variable, through poor sampling, anisotropy, or refraction. Where no core was available, representative samples were selected from other boreholes penetrating similar rock units, or cuttings were used with cell techniques giving less accurate determinations. Heat flow was calculated in most instances by combining average conductivities with a least-squares

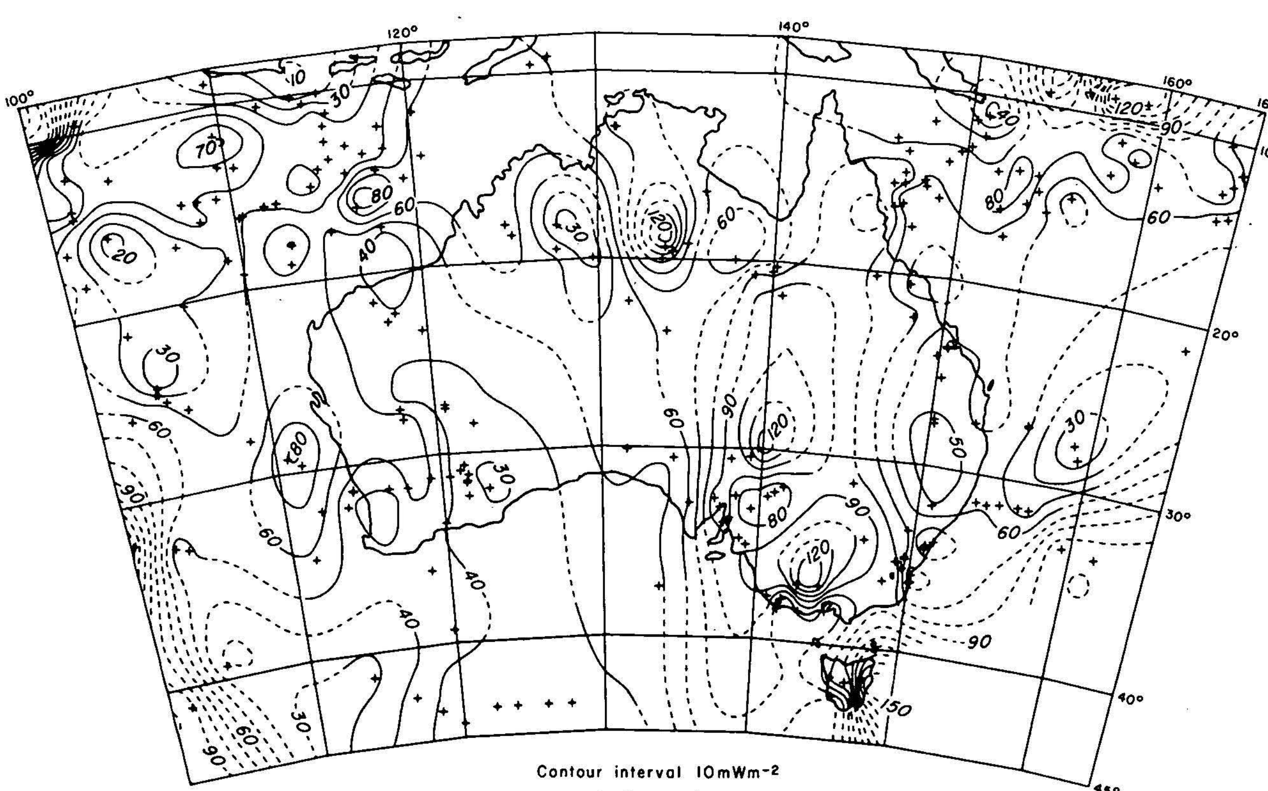

Figure 1. Heat-flow for the Australian region. Contours derived from a 1° grid of values.

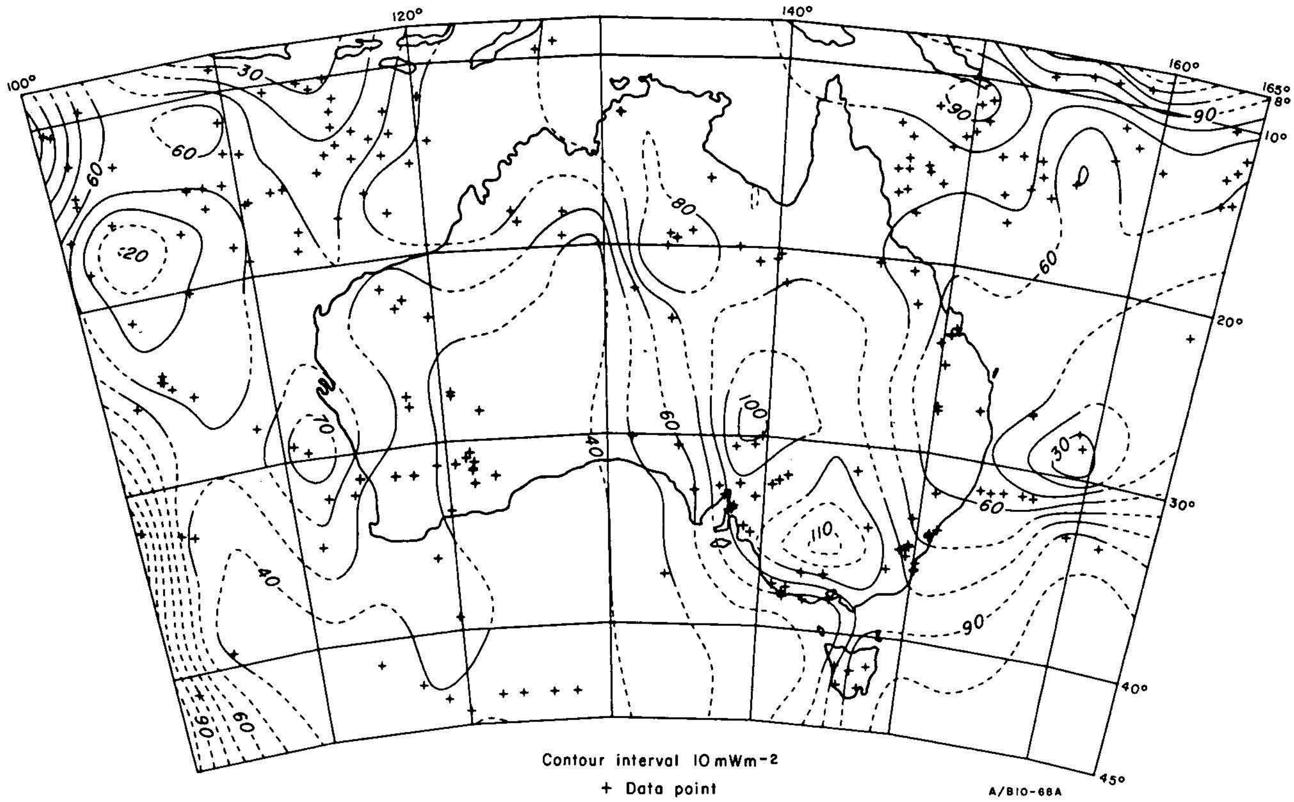

Figure 2. Heat-flow for the Australian region. Contours derived from a 3° grid of values.

thermal gradient over the full length of the hole. Occasionally several linear segments of the thermal gradient were identified, and heat flow was calculated for each. More recently, individual determinations of thermal conductivity have been combined with geothermal gradients calculated for appropriate 10 m intervals to give heat flow as a function of depth. Bullard-type reductions have been used primarily to verify results obtained with conventional techniques, in boreholes having complex lithologies.

Contour maps

The computer techniques developed by Murray (1977) were used to produce the contour maps in Figures 1 and 2. These techniques involve the generation of a grid of heat-flow values over the whole map area, and the subsequent production of a minimum curvature surface to fit the grid. Where there were no observed values within a grid square, the grid values were determined by extrapolation from nearby observations. For the map in Figure 1 equal weights were assigned to all data with a grid spacing of one degree, in an attempt to preserve the integrity of individual determinations. There is no obvious discontinuity between marine and land values and contours are generally smooth. Only in Tasmania and the Snowy Mountains is there any evidence for short-period anomalies unrelated to surficial deposits of uranium.

The contour map in Figure 2 was produced using a 3° grid, after weighting the data for reliability. Data were classified as excellent, good, fair, or poor, and weights were assigned from 4 to 1 respectively. Marine data were not assessed in detail and consequently all were assigned equivalent lowest weight. The previously noted short-period anomalies are significantly reduced, but the general character of the contour map is unchanged. Regional trends are obvious, with low values in the west (corresponding to the Precambrian shield) increasing north and east into regions of greater surface radioactivity and more recent tectonism.

Low values are observed on the recharge margins of the Great Artesian Basin; regional values may need correction in these areas. A corresponding discharge zone of high heat flow is apparent in South Australia north of Spencer Gulf. Hot springs in the area (e.g. Paralana) are probably associated with outflow from the Great Artesian Basin, but recent tectonism may also cause thermal anomalies. Values in the centre of the Basin should not be greatly affected, since any flow is nearly horizontal with low velocities (of the order of 1 m/yr) and negligible vertical component.

Generally high values are observed in regions of recent volcanism in Queensland and western Victoria, but there are no indications of near-surface geothermal phenomena. The most recent eruptions occurred at Mount Gambier (Sheard, 1978), but local anomalies may be dominated in this region by water movements in the Otway Basin.

Low values can be expected in regions of Quaternary glaciation; surface temperatures increase rapidly with glacier retreat and transient perturbations may persist, causing a decrease in the geothermal gradient (Crain, 1968). However, values previously reported for Tasmania and the Snowy Mountains are consistently above the world average (60 mW m^{-2}). No corrections have been applied for climatic changes and consequently the contours shown represent minimum values. Climatic corrections would result in significant positive anomalies unrelated to surface geology, particularly in the southeast where surface temperatures may have increased by 10°C following glacier retreat about 15 000 B.P. To resolve uncertainties in the data for this region, observations are required in boreholes at depths greater than 1000 m (Beck, 1977).

Discussion

Apart from the marine data in the Coral Sea, east of the north Queensland coast, offshore values are usually low, particularly near the subduction zone associated with the Java Trench, where several values less than 40 mW m^{-2} have been obtained. Transitions at the edge of the continent are not manifest in the contour pattern, except between Spencer Gulf and Tasmania, where the land values are higher than average and there is an almost complete absence of marine observations. The oceanic lithosphere close to the margin is usually older than 50 m.y. (except in the Coral Sea where the ages are thought to range from 30-35 m.y.) and the value of heat flow predicted from most thermal models would be between 30 and 60 mW m^{-2} (Sclater & Francheteau, 1970). On land the bulk of the continental crust is older than 250 m.y.; west of 135°E it is mostly older than 1000 m.y. and heat-flow values would therefore be expected to be in the range 35-55 mW m^{-2} (Polyak & Smirnov, 1968). Both oceanic and continental crusts are comparatively old in the Australian region and the heat flow/age curves are nearly flat for both types of crust, giving fortuitous agreement in heat-flow values.

Heat flow in eastern Australia generally exceeds 60 mW m^{-2}. Howard & Sass (1964) noted a general correlation with the age of the basement; they infer that the Precambrian Shield extends under all regions of low heat flow. However, high values may also be caused by (1) higher concentrations of radioactive materials in the crust—e.g. near Tennant Creek (20°S, 134°E), (2) large-scale vertical movements of heated water in sedimentary basins—e.g. along outflow margins of the Great Artesian Basin (30°S, 139°E) and (3) thermal events in areas of recent tectonism and/or volcanic activity—e.g. Southwest Victoria.

Correlations

Since most of the measured heat flow is generated by the decay of radioactive isotopes in crustal rocks (Roy & others, 1968) it is expected that correlations will exist between surface heat flow and factors such as the age, thickness, and type of crustal material present.

One of the clearest correlations in Australian data is that between heat flow and P-wave travel-time residuals. Seismic data were obtained from Everingham (1969), Everingham & Gregson (1971), Cleary (1967), and Cleary & others (1972). Residuals were available for 19 stations where heat-flow values had been determined close by. These results are plotted in Figure 3, and show a marked positive correlation. The high heat-flow values at WRA and PNA can be excluded since they probably result from high levels of radioactivity in the crust, and the effects of outflow from the Great Artesian Basin, respectively.

Correlations between crustal thickness and heat flow are not well established for Australian data. Figure 4 shows a heat flow/crustal thickness plot for the observations with the greatest weight. The crustal thickness data are taken from the work of Dooley (1972), Wellman (1976) and Finlayson & others (1978), and are rounded off to the nearest five kilometres. Although there is no obvious correlation between heat flow and crustal thickness, the available data, particularly those on crustal thickness, are not yet good enough to dismiss the possibility.

A further correlation is noted between heat flow and upper mantle P-wave velocity. Figure 4 shows the values for six regions of the Australian continent where the P_n velocity is reasonably well known. These values were taken from Doyle & Everingham (1964), Underwood (1969), Mathur (1974), Collins (1978), and Collins, Drummond, & Finlayson (pers. comms.). The negative correlation probably reflects the temperature derivative of the P-wave velocity. In regions of low heat flow, mantle rocks are expected to be relatively cool with comparatively high velocities, while lower velocities are expected in regions of higher heat flow. Comparing heat flow in Western Australia and Eastern Australia temperature differences of 600°C may exist

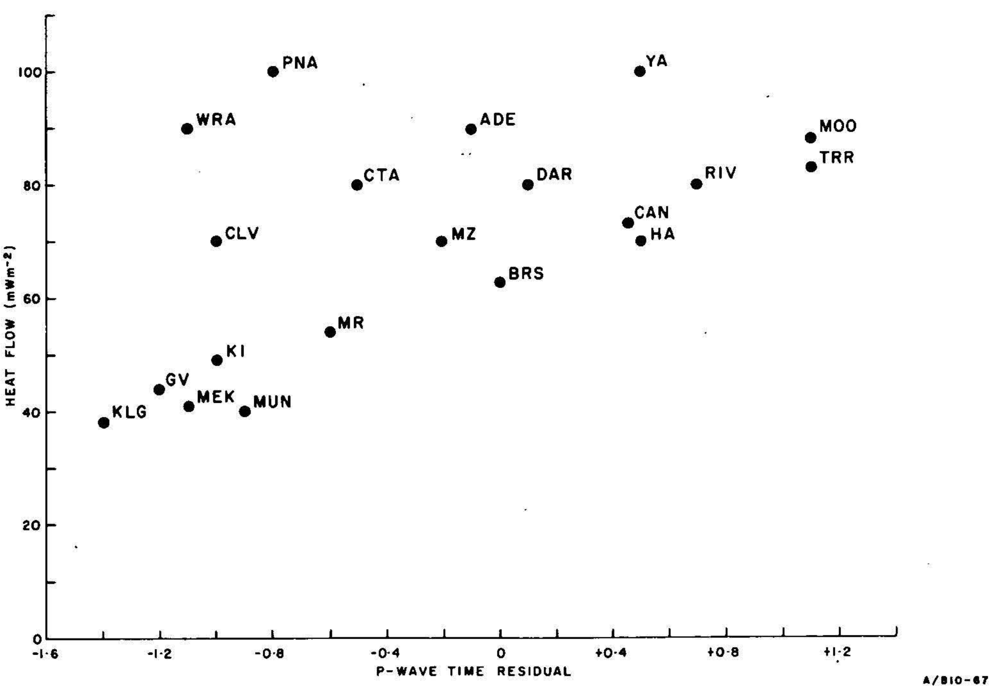

Figure 3. P-wave time residual versus heat flow. The station codes are listed in references quoted in the text.

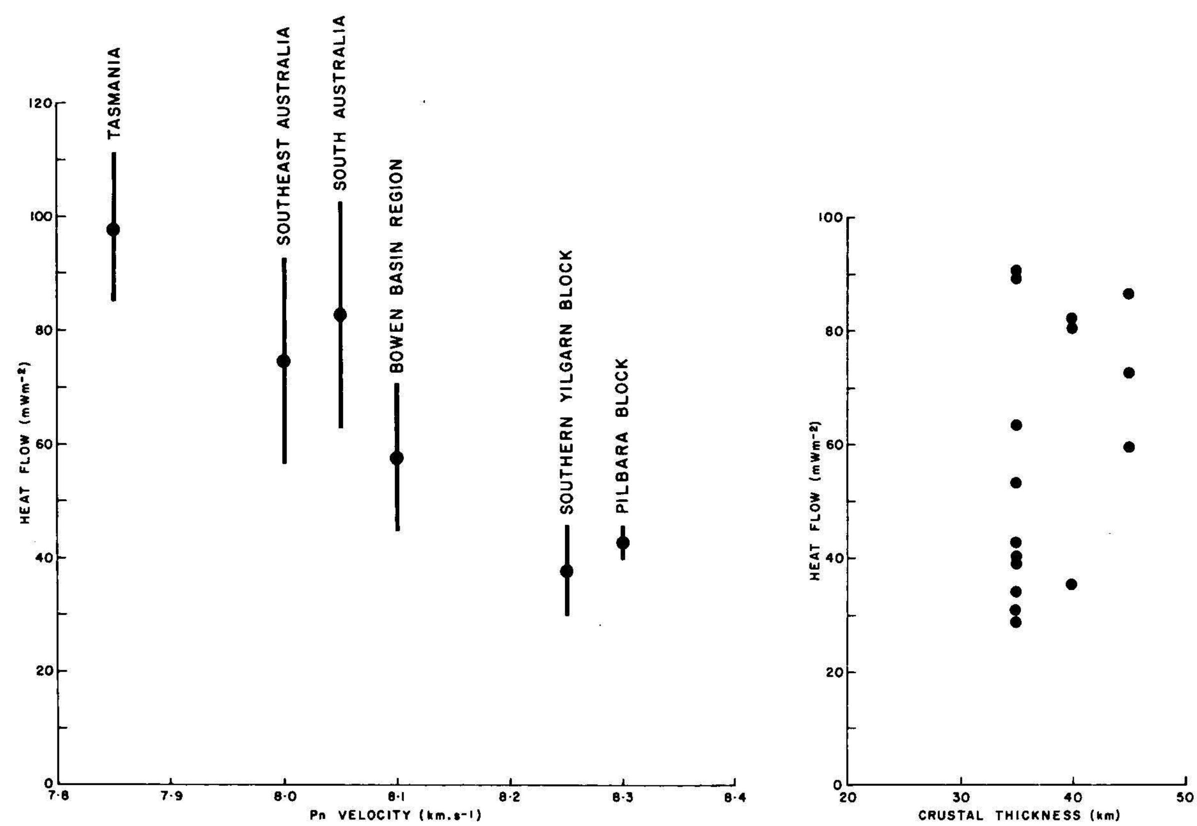

Figure 4. Pn velocity and crustal thickness versus heat flow. The bars on 6a indicate one standard deviation.

at the base of the crust (Sass & others, 1976). Consequently observed variations in velocity of 0.4 km/s are fully explained (derivatives $\delta V/\delta T \approx -0.0005$; Anderson and others, 1968) with no change in mantle chemistry.

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Appendix

New results from BMR surveys

This appendix contains details of the new results presented in this paper. Thermal conductivities were determined using standard divided-bar techniques (Birch, 1950; Cull & Sparksman, 1977). Where only drill cuttings were available the cell technique described by Sass & others (1971) was used to contain such fragments in the divided bar. However, it has been noted previously (Cull & Sparksman 1977) that for the BMR apparatus, the cell technique gives results which require a correction of —20 percent, because the divided bar is calibrated with solid core, which has different side losses to those of the cell. The correction was applied to all results obtained with the cell technique. The thermal logs are shown in Figures 5-7.

North West Australia

Noonkanbah. This area has been drilled extensively as a coal prospect. Temperature/depth profiles were recorded in two abandoned holes penetrating the Permian Liveringa Formation. Thermal conductivities were determined on thirteen core samples from the stratigraphic boreholes Noonkanbah Nos. 1 and 2. Since the temperature/depth profiles from both ESSO 15 and 22 were similar, an average thermal gradient was adopted. This was combined with the mean value of thermal conductivity, and heat flow was found to be 67.2 ± 3.5 mW m^{-2}*.

Billiluna No. 3. Core was available at a depth of 65 m from borehole BIL No. 3 in the Upper Devonian Knobby Sandstone of the Billiluna Group. A single determination of thermal conductivity was combined with the local geothermal gradient. The heat flow was calculated to be 46.4 mW m^{-2}, but conductivity sampling errors are unknown. Cuttings were also available in this borehole and the cell technique was used to obtain an average thermal conductivity for the interval with linear geothermal gradient (40-70 m). Heat flow was calculated as 44.6 ± 2.3 mW m^{-2} — in close agreement with the result from solid core. A geometric mean value of 45.8 ± 2.3 mW m^{-2} has been adopted as a representative value of Q for this location.

Bannerman No. 1. The second hole near Billiluna (BAN No. 1) penetrated the Permian Lake Gregory Beds to a depth of 65 m. Considerable difficulties were experienced in measuring conductivities with the cell technique, because of the tendency of the clays to swell during saturation and escape confinement. Furthermore, since the original *in situ* porosity was unknown, a value of 10 percent was assumed for data reduction. A corrected heat flow of 58.7 ± 5.7 mW m^{-2} was obtained using an average thermal conductivity from three samples at Cornish No. 3 bore. This result is about 30 percent higher than the value obtained at Billiluna No. 3, and since there were considerable problems with the conductivity measurements, the result at this site is considered unreliable.

Tanami. Drill cuttings were obtained at depths from 50 m to 80 m for two separate water bores in the Lower Proterozoic Tanami Complex. Four determinations of thermal conductivity were averaged and combined with the linear segment of geothermal gradient in the depth interval 50-95 m. The corrected heat flow was calculated to be 34.4 ± 6.2 mW m^{-2}. If the single low value of conductivity is omitted, heat flow is increased to 39.1 ± 3.0 mW m^{-2}.

Halls Creek. Only two samples of cuttings were available for the Lower Proterozoic Halls Creek Group. The cuttings were obtained at 10 m and 60 m for a single hole (76/1A) close to the holes which were logged for temperature. No control was available for estimates of original porosity and a value of 10 percent was assumed. Both samples gave similar, high results for thermal conductivity; this was expected from the very low thermal gradient (<11.0° C/km). Heat flow was calculated to be 25.6 mW m^{-2}. Although the statistical precision of the heat flow value is good the determination is considered unreliable because of the very low thermal gradient and the poor thermal conductivity sampling.

Yuendumu. Temperatures were measured in three waterbores penetrating Lower Proterozoic sediments. The bores were fully cased, but aquifer zones were perforated. No water had been pumped for twelve months, but the thermal gradients are obviously disturbed by water movements (Fig. 5); consequently heat-flow determinations are extremely difficult. Calculations can be confined to stable intervals, or alternatively a Bullard reduction used (Bullard, 1939; Cull & Sparksman, 1977). The Bullard reduction is a simple means of averaging data in the stable regions. An effective conductivity can be assigned to the disturbed interval, so that the Bullard plot is approximately linear. For Yuendumu the observations from No. 6 were used. Conductivities of 2.82, 4.0, and 2.24 Wm^{-1} K^{-1} were assumed for the intervals 0-90, 90-190, and 190-220 m respectively, resulting in a mean heat flow of 55.8 ± 3.1 mW m^{-2}. The conductivities for the upper and lower section are reasonably well determined, but that adopted for the centre interval was chosen to make the Bullard reduction linear. For the purposes of this reduction it was necessary to adopt a value for the surface temperature; a figure of 28.0°C was obtained by adding 3°C to the mean annual air temperature (Howard & Sass, 1964; Cull & Sparksman, 1977).

St George Range. Temperature data from the St George Range oil well were obtained from bottom-hole temperatures taken during pauses in drilling. These are shown in Figure 5, with the conductivity data which were determined from core samples distributed over the full length of the hole. If the mean thermal conductivity in each stratigraphic unit is combined with the apparent geothermal gradient, the average heat flow is calculated to be 62.7 ± 4.0 mW m^{-2} (one anomalous value was removed). This value agrees well with the Bullard reduction value of 64.9 ± 4.0 mW m^{-2}, and leads to an adopted value of 63.8 ± 4.0 mW m^{-2} for heat flow at St George Range.

Pilbara Region

Wittenoom 47A. This hole, which was logged to the extent of cable available (300 m), was drilled into the Proterozoic Brockman Iron Formation of the Hamersley Group. The temperature/depth plot is made up of four distinct segments. Cores were available in each segment, and separate calculations were made for heat flow. The data result in a weighted mean of 40.6 ± 2.9 mW m^{-2}.

Millstream No. 9. The Millstream hole was drilled as a waterbore penetrating Pleistocene sediments associated with the Fortescue River. These lay above Proterozoic Wittenoom Dolomite, which is part of the Hamersley Group. Figure 6 shows the temperature/depth plot and the conductivity results determined from core samples. The curve has been divided into four segments for seperate determinations of heat flow. These data, taken in conjunction with results from Wittenoom 47A suggest that Q possibly increases with depth. Low near-surface values may be caused either by secular changes in climate, or possible groundwater movements—both of which affect the thermal gradient. The weighted mean of 40.5 ± 1.3 mW m^{-2} is nevertheless adopted as representative of the region.

Tom Price G623. The hole at Tom Price was drilled into the Tertiary hematite deposits which lie above the Brockman Iron Formation. The temperature data are not as good as the Millstream or Wittenoom holes, because there was no water in the hole and times required for probe equilibration were erratic. However the stratigraphic column, which consists of shale and hematite bands, is well known. Conductivities were obtained from cores of one shale sample (3.0 Wm^{-1} K^{-1}) and three hematite samples (6.51, 8.20 and 10.88 Wm^{-1} K^{-1}). Because the core that yielded the 6.51 result was slightly weathered a value of 9.5 Wm^{-1} K^{-1} was adopted for the conductivity of hematite. Figure 6 shows the temperature and stratigraphic logs. These data were combined to obtain a Bullard reduc-

*** All errors quoted here refer to the standard deviation of the mean unless otherwise specified.**

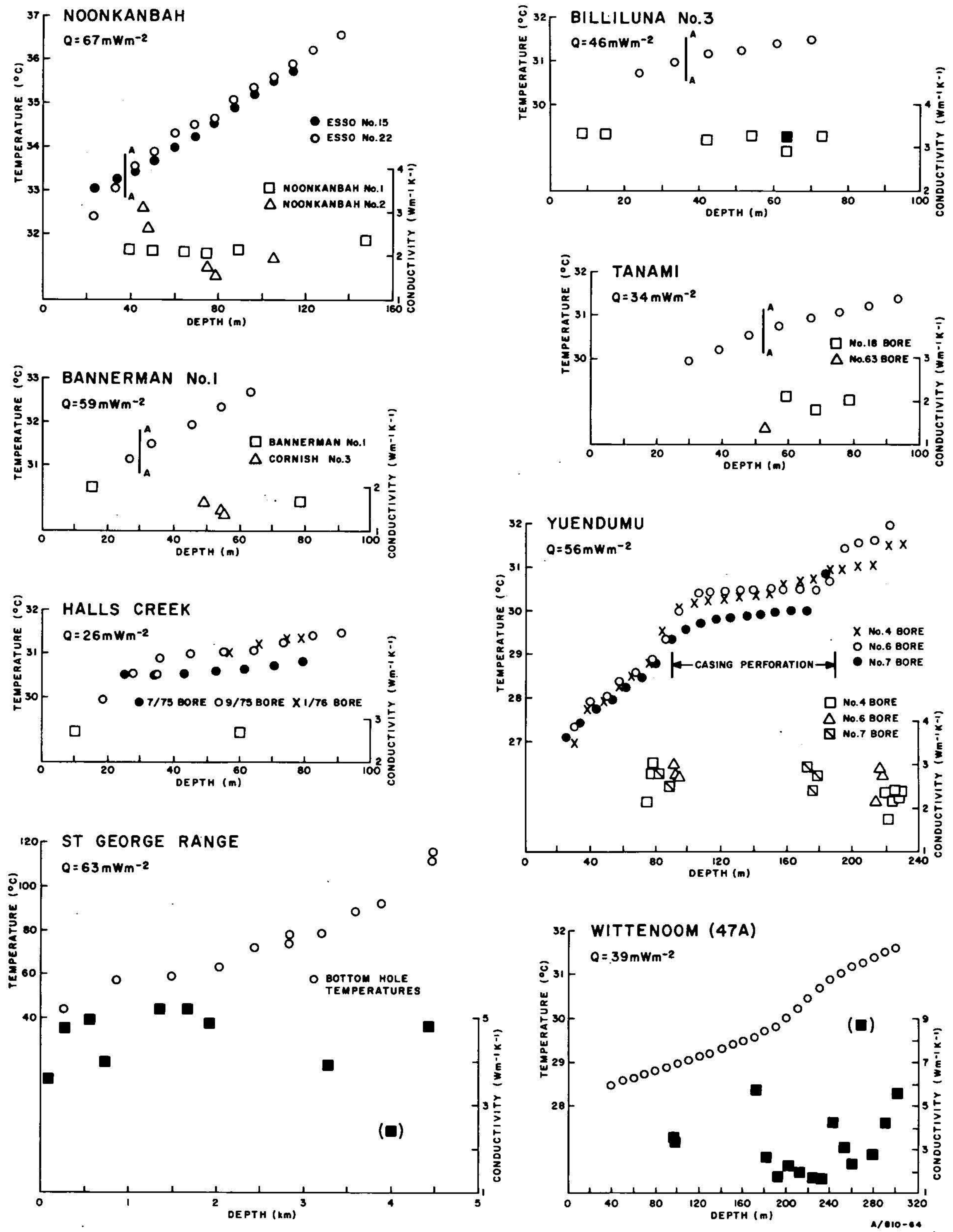

Figure 5. Heat-flow data from boreholes.

The graphs show the actual temperature measurements down the holes, except for St George Range where bottom-hole temperatures measured during pauses in the drilling are plotted. Points on the left of the vertical AA lines were not used to compute the temperature gradients. Where the conductivity values are plotted with open symbols the cell technique has been used; solid symbols indicate the use of solid core. Conductivity results in parenthesis were not used to calculate the heat flow. At Yuendumu the casing perforation allowed water to flow between aquifers and hence perturb the temperature/depth plot between 90 and 190 m.

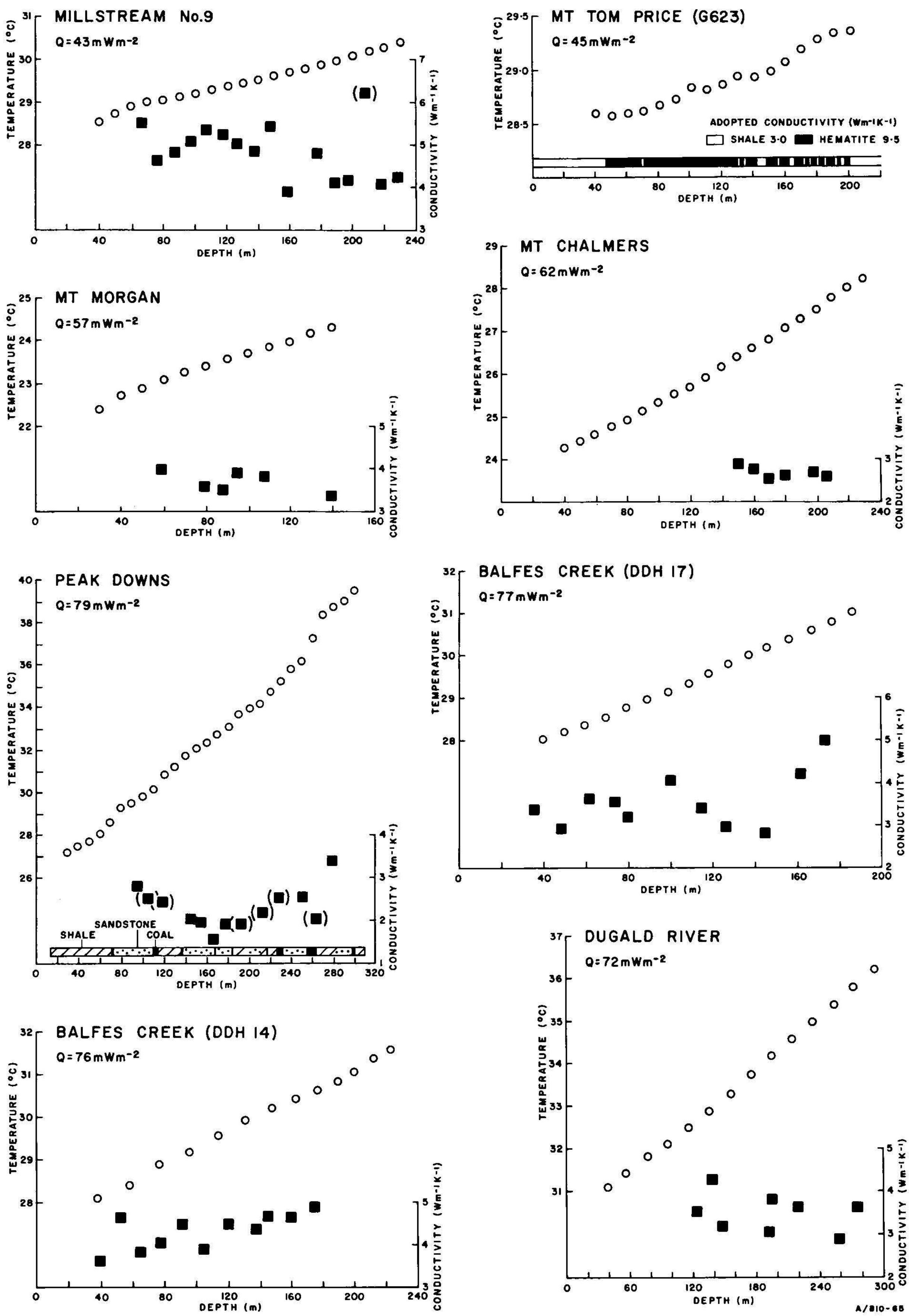

Figure 6. Heat-flow data from boreholes. Symbols are the same as for Figure 5.

tion from 50-190 m, and a value for Q of 45.0 ± 1.9 mW m^{-2} was computed.

North Queensland

Mount Morgan 63/3. Temperatures were measured to a depth of 140 m in a borehole 2 km northwest of the open-cut mine, penetrating rhyolite within the Middle Devonian Mount Morgan Tonalite. Curvature is evident in the temperature/depth plot to a depth of 60 m (Fig. 6). However, the gradient is essentially linear in the interval 60-140 m; a result of uniform thermal conductivities. No terrain corrections were applied for the effect of the open cut or for local topography (general relief <30 m). Heat flow was calculated for the linear segment (from 60-140 m) by combining individual determinations of thermal conductivity with apparent geothermal gradient over 10 m intervals. This procedure results in a value 56.8 ± 1.2 mW m^{-2}. The only previous result for this region was obtained 20 km northeast of Mount Morgan (Hyndman, 1967), where the apparent heat flow was 34.3 mW m^{-2}; corrections were subsequently applied to give a value of 54.3 mW m^{-2}. This higher value was considered more accurate and is confirmed by present results.

Mount Chalmers MW4. Temperatures were measured to a depth of 230 m in a single borehole situated in the lower Permian Berserker Beds. Topography is undulating, with relief less than 20 m. No terrain corrections were applied. Gradients appear to increase with depth (Fig. 6), and two linear segments can be identified. Core samples were obtained only for the deeper segments from 150-230 m. For 10 m intervals, apparent gradients were combined with individual determinations of conductivity. Heat flow was calculated to be 62.3 ± 0.9 mW m^{-2}. This result is consistent with the value 56.8 mW m^{-2} determined for Mount Morgan, approximately 50 km to the southwest.

Peak Downs Utah 22257. The Peak Downs Upper Permian coalfields have been extensively drilled. The stratigraphy is well known and full core is available. For most units thermal conductivity could be determined using normal divided-bar techniques. However for coal sequences, sampling is biased to well-consolidated core. Conductivity was measured for only one sample of coal using the cell technique to contain core chips (Sass & others 1971). Low-interval conductivities are indicated by the very high thermal gradients observed in the coal beds. Heat flow can be determined from the apparent gradient only if conductivities are representative of interval lithology. For this reason data close to coal seams were omitted. Using the remaining data, heat flow is calculated to be 79.2 ± 5.6 mW m^{-2}. Thermal conductivities are most consistent in the interval 120-200 m. As a result the temperature/depth profile is nearly linear in this segment. Heat flow can be determined from an average thermal conductivity and a least squares temperature gradient. This results in a value of 76.3 ± 4.1 mW m^{-2}. Because of the complex lithology a Bullard plot was used to confirm the above result. With this technique all data are used and resistivities are assigned to each unit identified on the statigraphic logs. In practice conductivities are assumed to be constant in each interval, except for coal segments. A least-squares fit (Fig. 6) results in a value of 80.7 ± 1.9 mW m^{-2} for apparent heat flow. Because of bias to well-consolidated core this value is probably higher than actual. Results have been reported previously for Collinsville, 175 km north of the present location (Hyndman, 1967). Heat flow was calculated to be 52.7 ± mW m^{-2}, substantially less than the preferred value of 79.2 ± 5.6 mW m^{-2} obtained at the present site.

Balfes Creek Le Nickel. Temperatures were measured in two closely spaced exploration boreholes, DDH14 and DDH17, penetrating rhyolite in the Cambrian/Ordovician Mount Windsor Volcanics. Temperature/depth profiles are similar down to a depth of 150 m (Fig. 6), but then diverge. One linear segment is sufficient to describe data from DDH17, but three can be identified in DDH14. Intervals of low gradient are generally characterised by increased thermal conductivities. In both cases apparent gradients for 10 m intervals were combined with local values of thermal conductivity, to give heat flow as a function of depth. For DDH14 there appears to be a significant decrease in heat flow in the interval 150-200 m. Low gradients are observed in this interval, and it is possible that the ore body (not apparent in DDH17) perturbs the geothermal field. Heat flow is calculated to be 75.7 ± 4.0 mW m^{-2} and 77.4 ± 4.7 mW m^{-2} for DDH14 and DDH17 respectively.

Dugald River CRA. Dugald River is a metallogenic shale prospect located in Lower Proterozoic sediments. Temperatures were measured in a steeply dipping borehole (DR40) four weeks after completion of drilling. No corrections were applied to the data (except for calculations of vertical depth) and absolute values may be in error by 2-3 percent (Jaeger, 1961). However any such error is negligible in calculations of gradient. Although there is a scatter in thermal conductivities (Fig. 6), the geothermal gradient is remarkably linear, with only one major fluctuation, at 90 m. Heat flow was calculated to be 72.0 ± 2.9 mW m^{-2}.

Mt Dore Amoco. Temperatures were measured in a single vertical borehole penetrating metamorphosed Lower Proterozoic sediments and porphyritic granites to a depth of 195 m. Three segments with constant gradient were observed on the temperature/depth profile (Fig. 7). Representative cores were taken for each segment. Thermal conductivities were combined with apparent gradient over 10 m intervals. Heat flow was calculated to be 98.2 ± 4.1 mW m^{-2} (4 samples).

Miscellaneous

Canberra: BMR 145. This hole penetrates Silurian limestone to a depth of 260 m. Although there is curvature in the geothermal gradient (Fig. 7) there are corresponding changes in thermal conductivity. Heat flow is calculated as a function of depth by combining apparent thermal gradients and local conductivities. A systematic increase in heat flow is observed to depths of at least 120 m, indicating a long-term variation in surface temperature. For depths greater than 120 m, heat flow is calculated at 72.5 ± 2.7 mW m^{-2}. Four linear segments are apparent in the temperature/depth profile. Using average conductivities and least-square gradients in the intervals 70-120 m, 120-180 m, 190-200 m and 230-260 m heat flows are calculated to be 48.6, 70.24, 73.78, and 74.19 mW m^{-2} respectively. The systematic increase with depth indicates surface temperature perturbations consistent with glacier retreat in the Snowy Mountains (Bowler, 1976). Corrections of up to 30 percent may therefore be required in the final solution. The value of 72.5 ± 2.7 mW m^{-2} reported here is lower than results reported previously for the Canberra region: 86.1 ± 1.6 mW m^{-2} at Mount Stromlo, and 100.3 ± 3.4 mW m^{-2} at Captains Flat (Sass & others, 1976).

Woodlawn: Jododex CE 12. Temperatures were measured in a steeply dipping exploration borehole penetrating Silurian rhyolite to total actual depth of 260 m. Core was available for thermal conductivity measurements at approximately 20 m intervals. Although curvature is evident in the temperature/depth profile (Fig. 7) there is little corresponding change in thermal conductivities. A significant systematic increase is noted for heat flow calculated as a function of depth, indicating climatic perturbation in the region consistent with results for Canberra. The apparent heat flow is calculated to be 59.6 ± 1.5 mW m^{-2}—significantly less than values of 79.4 mW m^{-2} previously reported (Sass & others, 1976). If climatic corrections are made, the values could be increased by up to 30 percent.

Ardlethan: Tin mines. Temperatures were measured in three exploration holes drilled in Silurian granites near the open-cut tin mines at Ardlethan. Linear gradients were obtained for the intervals 50-140 m—average of 3 holes with 60 m spacing—and 150-200 m from 1 hole (Fig. 7). No core was available and drill cuttings were used to measure thermal conductivity with cell techniques. Thermal

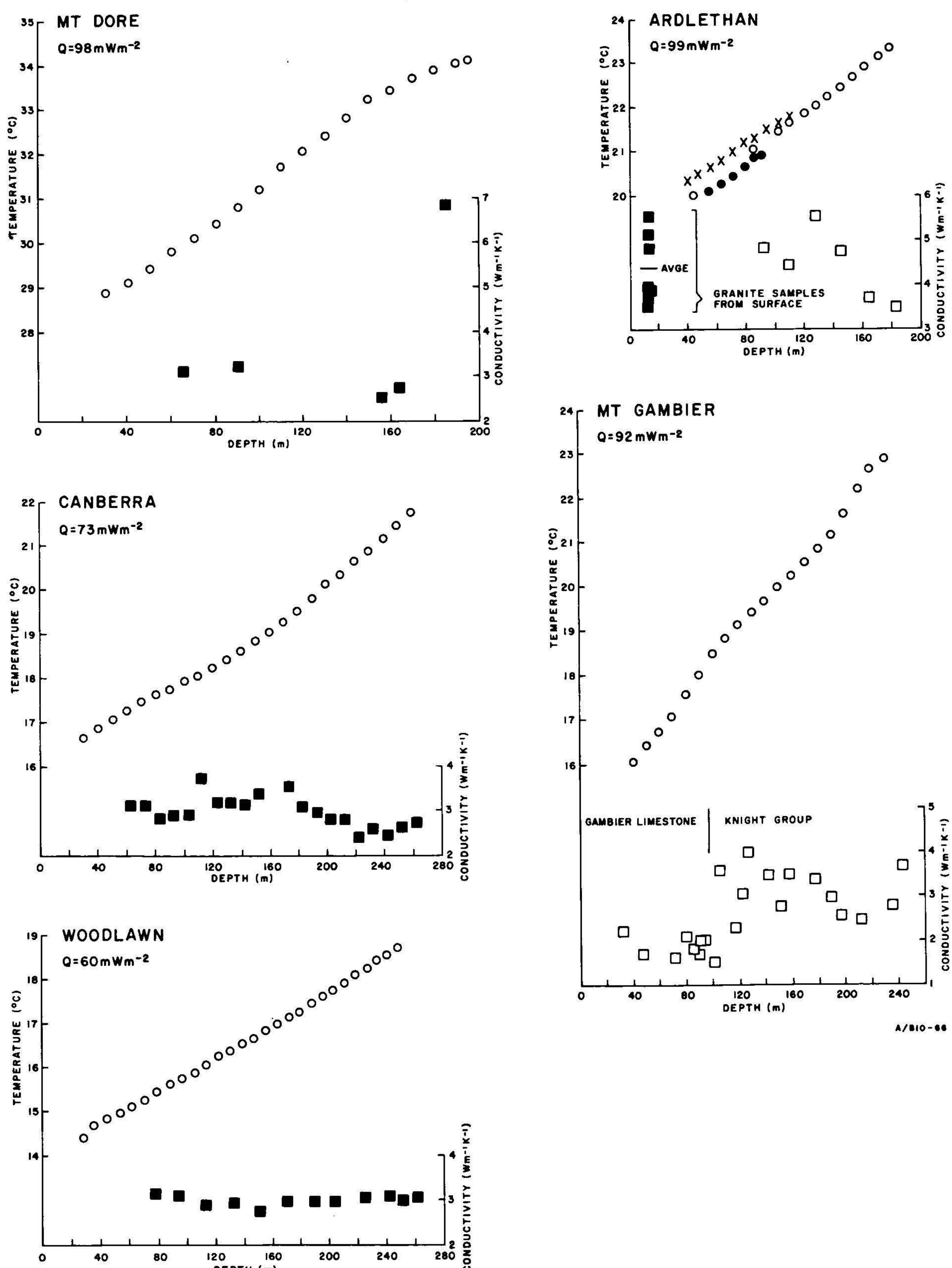

Figure 7. Heat-flow data from boreholes. Symbols are the same as for Figure 5.

conductivities in the deep hole were combined with the temperature gradient in successive 10 m depth intervals to give a heat flow value 105.4 ± 4.4 mW m^{-2}. In addition to drill cuttings, representative whole rock samples were obtained from the open cut. The average temperature gradient from each hole was then combined with the mean rock conductivity. Heat flow was determined as 92.9 ± 5.9 mW m^{-2}, which is in reasonable agreement with the previous value obtained from drill cuttings. An average value 99.2 mW m^{-2} is adopted for the region.

Mount Gambier: BMR PEN 31. Mt Gambier is the centre of the most recent (4500 B.P.) volcanic eruption in Australia (Sheard, 1978), and it is possible that large quantities of heat still remain (Cull, 1978b). To assess prospects for geothermal energy, a 250 m hole was drilled within 15 km of the crater. The hole was fully cased and temperatures were measured 8 months after drilling. The hole penetrates Gambier Limestone and Tertiary Knight Group sands (Fig. 7). Core samples were taken for measurements of thermal conductivity, but all samples were soft and cell techniques were required. Heat flow was calculated as a function of depth by calculating temperature gradients for each 10 m interval in which thermal conductivities were determined. There is considerable scatter in the results, but no systematic variation is apparent. A mean value of 91.8 ± 3.0 mW m^{-2} is adopted for the region.

Evidence of an exhalative origin for deposits of the Cobar district, New South Wales

D. F. Sangster[1]

Evidence is presented to show that a strong mineralogical/chemical zoning exists in seven deposits of the Cobar-Nymagee area. Characteristically, the within-deposit zoning, perpendicular to bedding, consists of a siliceous chalcopyrite-pyrrhotite eastern side with diffuse contacts against adjacent siltstone-shale host rocks, and a relatively massive sulphide, banded, pyrite-sphalerite-galena western side with sharp contacts against host rocks. Features such as these are typical of those in exhalative deposits in volcanic terrain and are taken here to indicate a similar origin in this essentially non-volcanic environment.

The deposits are contained in distal turbidite facies of the Devonian Cobar Supergroup, deposited in a meridional trough bounded on its eastern flank by a possible penecontemporaneous growth fault separating the trough from an adjacent shelf area on which were deposited shallow-water marine sediments and terrestrial and marine volcanics. This volcanism and the Cobar sedimentary-exhalative deposits may be related through rifting in the area which produced the proposed growth faults and the subsequent Cobar Trough.

The deposits, now in 20° discordancy to bedding, are considered to have been transposed into the prominent regional cleavage during post-ore deformation.

Using the syn-sedimentary exhalative concept, two mineralised horizons and a possible tight syncline may be recognised in the CSA mine.

Stretching 120 km from the Nymagee deposit in the south to Elura in the north, the Cobar area mineral belt (Fig. 1) has produced, to 1973, approximately 11.7 million tonnes of ore averaging 1.7% Cu, 3.25 g/t Au, and 10 g/t Ag, as well as lesser amounts of lead and zinc (Brooke, 1975, p. 687). In 1975, published 'ore potential' for the CSA deposit was 31 Mt, for the Chesney 11 Mt, and for the Gladstone 2 Mt (BH South Ltd, Annual Report 1975). Ore reserves for the, as yet undeveloped, Elura deposit are 27 Mt at 8.5% Zn, 5.5% Pb, and 137 g/t Ag (Mineralogical Magazine, 1977). This results in a total of over 80 Mt of ore and ranks the Cobar area as a significant mineral belt.

The orebodies exhibit a marked structural control and, over the 100 years of mining in the area, authors have ascribed an epigenetic, hydrothermal origin to the ores. A summary of previous opinions on ore genesis at Cobar was presented by Rayner (1969, p. 48) who himself concluded the ores were epigenetic, high-temperature hydrothermal deposits.

Previous workers had suggested the deposits were replacements controlled by structures in fractured slaty rocks (Thompson, 1953) or veins in axial plane cleavage in a suitable structural environment (Kappelle, 1970). Recently, however, Brooke (1976, p. 29) admitted that 'there is some suspicion of an original sedimentary origin' for the deposits in the immediate Cobar area. Similarly, Suppel (1976, p. 31) suggested that the Nymagee lodes 'could have been remobilised out of bedding into the cleavage during shearing'. Neither author, however, elaborated on the concept nor presented evidence in support of a syn-sedimentary origin. Recently, Kemezys (1978, p. 105) considered the Cobar deposits 'could be examples of an area in which hot metal-bearing fluids permeated the strata but neither they nor the underlying hot magmas reached the surface', i.e., a classical sub-surface hydrothermal deposit.

1. Geological Survey of Canada, 601 Booth Street, Ottawa, Canada KIA OE8. Dr Sangster was in Australia in 1977-78 on an exchange agreement between BMR and the Geological Survey of Canada.

In the present brief review, the author will present evidence for a syn-sedimentary exhalative origin for deposits of the Cobar-Nymagee area.

Stratigraphy

Details of the geology in the Cobar area may be found in numerous publications including, Brooke (1976), Pogson & others (1976), Gilligan (1974), Brunker (1969, 1973), Baker, Schmidt, & Sherwin (1975), and Pogson & Felton (1977). Obviously the area is still under investigation and inasmuch as the regional stratigraphy has not yet been fully determined, the author has adopted the scheme put forward by Pogson & Felton (1977). For purposes of this discussion, the geology of the area will be considered in terms of two broad units: (1) basement; and (2) cover rocks consisting of the Cobar Supergroup and its equivalents.

Basement

In the area in question (Fig. 1), basement consists of intensely deformed Girilambone Beds (quartz-mica and quartz-albite schists), Ballast Beds (micaceous quartz arenite and phyllite), the Nymagee Igneous Complex (foliated and gneissic granite, minor migmatite), and the Erimeran Granite (coarse porphyritic granite). The latter is considered one of the youngest basement rocks and is thought to be Middle Silurian (Pogson & others, 1976, p. 32).

Cobar Supergroup

In the Cobar district, the Cobar Supergroup consists of the Nurri and Kenmure Groups occurring west of the main area of basement rock; the Kopje Group, an equivalent group seen to be lying on basement; and the Amphitheatre Formation.

Nurri Group

Pogson & Felton (1977) proposed the term Nurri Group to refer to the sequence in the immediate Cobar area and Kenmure Group in the Nymagee area. The two groups comprise the lower part of the Cobar Supergroup and interfinger northwest of Nymagee. Inasmuch

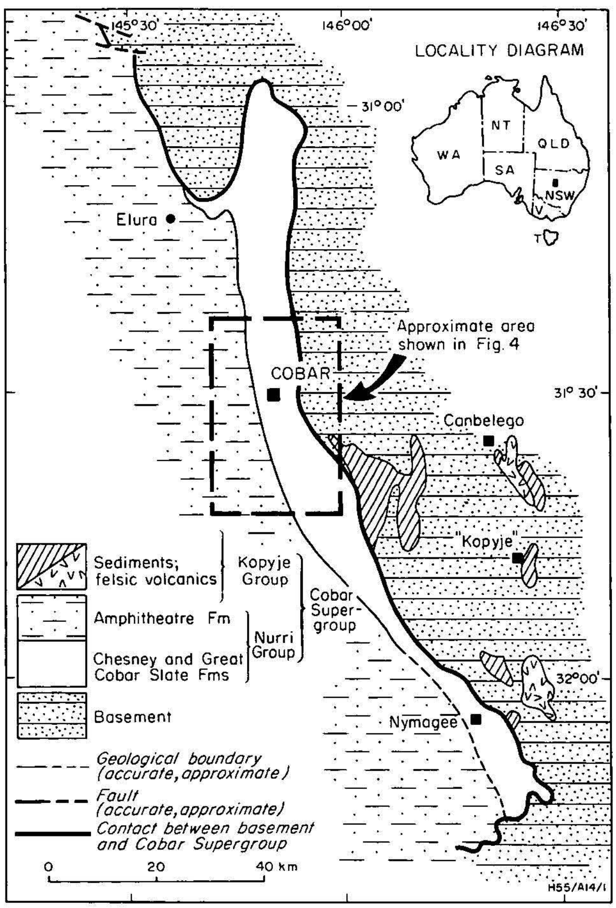

Figure 1. Generalised geology of the Cobar district. Simplified from Baker & others (1975), and Gilligan (1974).

as the Nurri Group contains a majority of the deposits considered in this discussion, it is briefly described here.

In the Cobar area, the Nurri Group consists of two formations—the Chesney Formation and the Great Cobar Slate.

Chesney Formation. This formation consists of sandstone, siltstone, shale and a basal conglomerate referred to as the Drysdale Conglomerate Member (Baker & others, 1975) in the area north of Cobar and as the Bee Conglomerate Member (Pogson & Felton, 1977) south of Cobar. These rocks are all regarded as having been part of a turbidite fan system. Pogson & Felton (1977) suggested the clastics of the Chesney Formation were derived from the east and northeast of Cobar. Equivalent units in the Nymagee area to the south were derived from sources to the west and south, mainly from the Erimeran Granite (Pogson & Felton, 1977).

Great Cobar Slate. This formation consists of shale and siltstone with a few sandstone beds and is regarded as a distal turbidite facies (Pogson & others, 1976). As with the Chesney Formation, source area appears to have been from the east and northeast (Pogson & Felton, 1977).

Amphitheatre Formation

Covering a large area to the west, southwest, and northwest of Cobar, the Amphitheatre Formation consists mainly of thinly bedded claystones, siltstone, fine to coarse-grained sandstone and quartzite, and a basal turbidite facies, the CSA Siltstone Member. In contrast to the underlying formation, the Amphitheatre is considered to be dominantly shallow water sediments derived from sources to the southwest, west and northwest (Pogson & Felton, 1977).

Kopje Group

In the area in question the Kopje Group commences with the sporadic development of a quartz lithic conglomerate to coarse arenite facies representing fluvial to shallow marine outwash fans. The remainder of the Group is represented by shallow marine to terrestrial siltstones and felsic volcanic rocks and flanked by carbonates representing possible atoll-like reef and back-reef facies (Pogson & Felton, 1977).

The Nurri, Kenmure, and Kopyje Groups contain similar Early Devonian macrofauna (Pogson & Felton, 1977). A schematic representation of stratigraphy and facies in these Groups is shown in Figure 2.

Structure

Basement structures are characterised by tight folding with well-developed slaty cleavage commonly deformed by a later NNW to north-trending crenulation cleavage. Except for the aureoles of contact metamorphism related to plutonic rocks in the Nymagee Complex and the Erimeran Granite, the grade of metamorphism in the basement is lower to middle greenschist facies (Pogson & others, 1976).

In the Cobar Supergroup, folding is generally described as broad (e.g. Kappelle, 1970, p. 81; Brooke, 1975, p. 685) but the same authors also describe the rocks as steeply west-dipping, which seems at variance with the concept of broad open folding. Near the centre of the district, the Cobar Supergroup 'forms the west limb of an anticline plunging 45°S. Average strike in the area is 165° with 50°-80° west dips. Minor overturning occurs, but east dips and facings are rarely seen.

'A strong regional cleavage, which is parallel to the axial plane of folding, strikes 155° and dips 75°-85°E. Movement has occurred in varying degrees sub-parallel to this cleavage, culminating in shears. Dominant stress relief has been east-block-north movement along the shears, and there is little cross faulting' (Brooke, 1976, p. 27).

Tectonics

Several authors (e.g. Kappelle, 1970; Pogson & others, 1976; Baker & others, 1975; Scheibner & Markham, 1976) refer to sediments of the Cobar Supergroup as having been deposited in a trough. Pogson & others (1976) refer to an 'extensional environment' and Baker & others (1975) consider the basin to have been a 'fault bounded trough'. Scheibner & Markham consider the Cobar sediments to have been deposited in a trough formed as a result of rifting during the Bowning tectonic stage (Silurian). Pogson & Felton (1977) further consider sedimentation of the Kopyje Group to have taken place in 'gently subsiding half grabens'. No reason is given for the abrupt change in sedimentary environment from the essentially terrestrial to shallow marine of the Kopyje Group on the east to the deeper water turbidites of the Chesney Formation and Great Cobar Slate to the west. Baker & others (1975), however, suggest the contact between basement and Nurri-Kenmure Groups is, at present, a fault. If such a fault were active during at least the lower part of the Nurri-Kenmure Group sedimentation, in the manner of a half-graben, then the presence of shallow marine to terre-

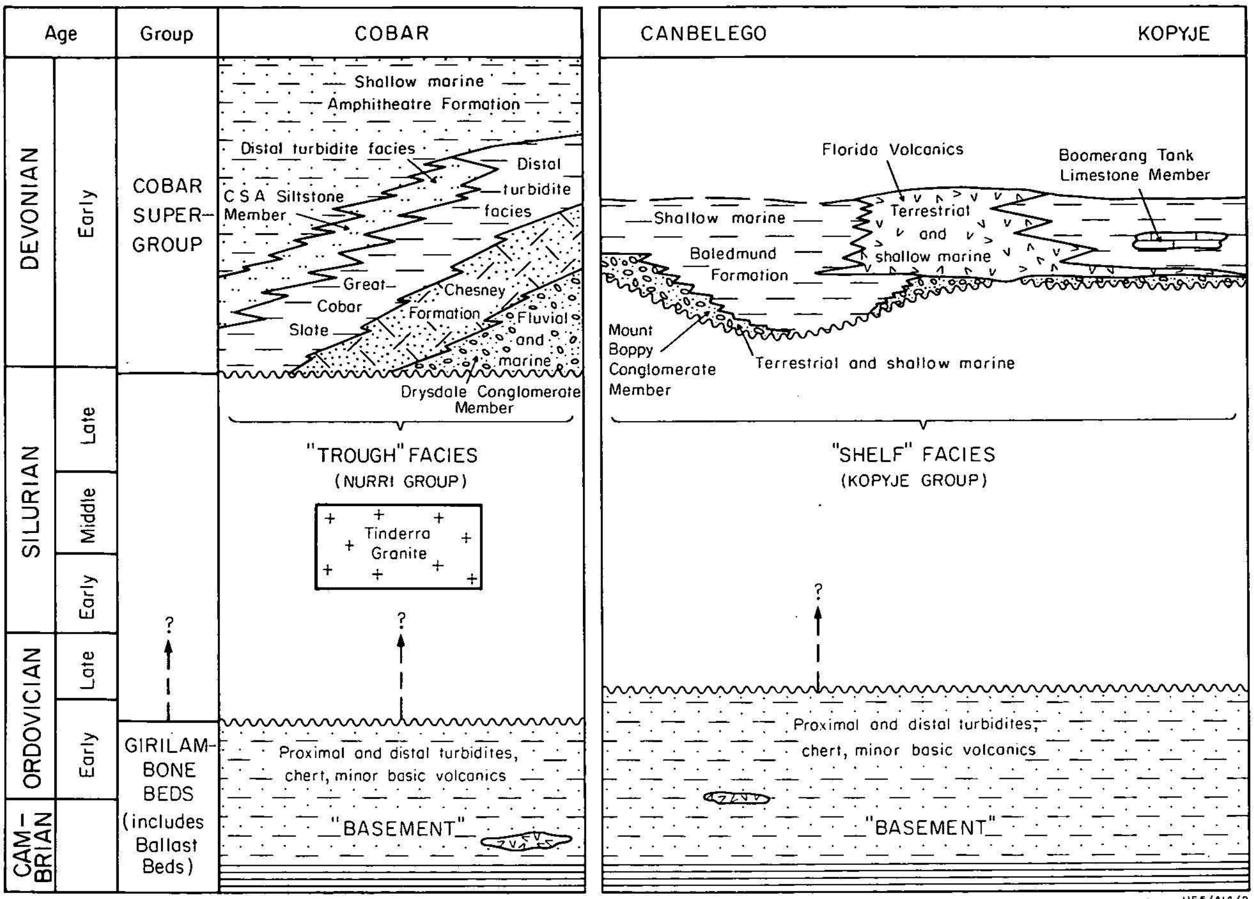

Figure 2. Schematic representation of stratigraphy and facies distribution, Cobar area. Modified slightly from Pogson & Felton (1977).

strial Kopyje Group on the upthrown east block could be reconciled with deeper water facies in the trough to the west.

The picture thus presented at the time when the Cobar Supergroup was being deposited is that of a deep meridional trough shallowing westward and bounded on the east against basement by a growth fault or group of faults (Fig. 3). Contemporaneously with deep-water sedimentation in the east-central portion of the trough, felsic terrestrial to submarine volcanism was taking place to the east and southeast. The sporadically developed conglomerate members, at the base of the distal turbidite Chesney Formation, may represent detritus from outwash fans shedding directly into the trough off the shallow water eastern shelf. Following infilling of the deeper portion of the trough by turbidites, the shallow-water Amphitheatre Formation transgressed eastward, marking the end of sedimentation in the trough.

Mineral deposits

Descriptions of individual properties in the immediate Cobar area (Fig. 4) may be found in references cited in Brooke (1975, 1976), and Gilligan (1974). The Nymagee deposit is briefly discussed by Suppel (1976). No published description of the Elura deposit is yet available and the author is guided by his own notes and observations collected during a brief visit to the property in January, 1978. The following general description of deposits in the immediate Cobar area is quoted from Brooke (1976):

'All known ore occurrences in the field lie within shear zones sub-parallel to the cleavage. . . . The oreshoots are similar in shape and attitude, but not necessarily in ore type or mineral composition. They strike slightly west of north, dip 70-85°E, and pitch steeply north. . . . Widths of fairly massive sulphides vary to over 30 m averaging perhaps 12 m. . . .

'The relationship observed between shoots across and along strike also exists vertically, but because of the common stockwork or vein-like nature of some shoots or their peripheries the exact structural pattern is not clear. Hence the size and grade of shoots is often determined more by mining methods and economics than by geological boundaries.

'The ore zones are roughly tabular, but in detail the oreshoots show considerable variation of dip and strike. Quartz and sulphide veins dominantly follow the cleavage, but also any other line of weakness—often sub-parallel fractures and sometimes the bedding. Occasional transverse veins 1 m wide are known. The bodies transgress the bedding, so have been subjected to structural control.

'There is also an apparent stratigraphic control. All the ore deposits occur within linear zones of higher magnetic intensity which roughly follow the stratigraphy, and three major orebodies and several small ones occur in the Great Cobar Slate adjacent to the Chesney Greywacke contact. However, this contact may also be regarded as a structural feature, as it represents a strong contrast in rock competency, and shearing has occurred along it. The emplacement of

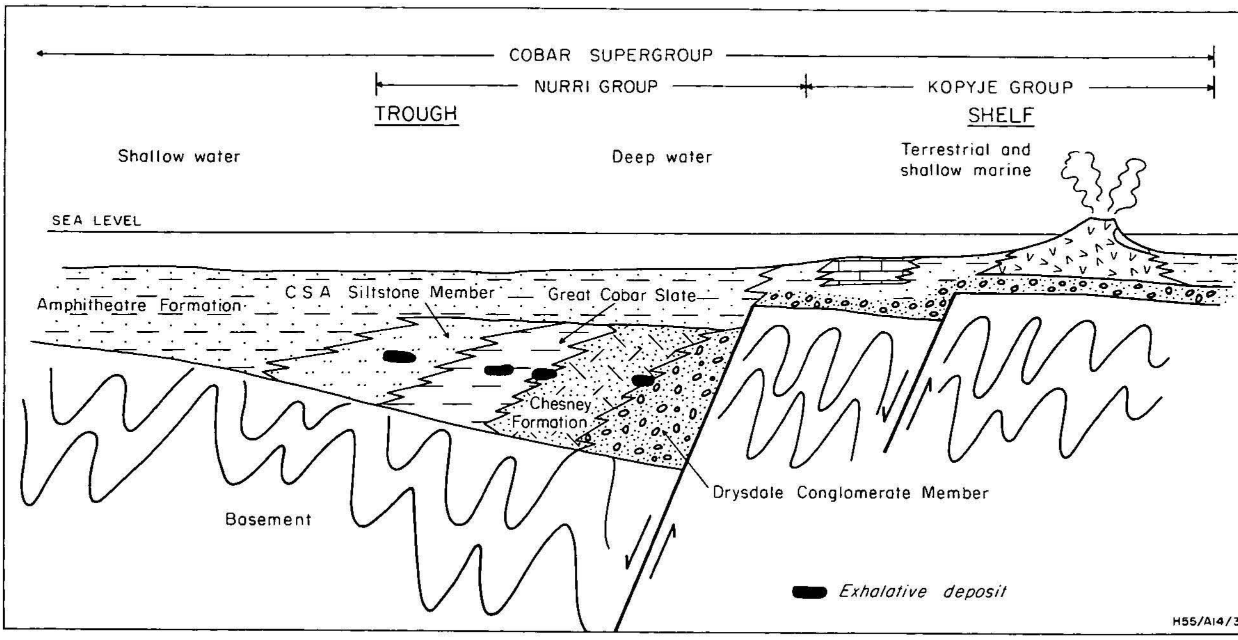

Figure 3. Schematic reconstruction of Cobar Supergroup and Kopyje Group sedimentary environments. Note the proposed growth fault between the two, and the possible time-equivalence of the four ore horizons in the Cobar district.

some of the other orebodies was possibly also related to contrasts in rock competency.

'Throughout the field, mineralization shows an obvious preference for finer grained sediments, but there is no satisfactory theory to explain the position of orebodies within the favourable structural or stratigraphic horizon.'

With the orebodies lying in the plane of cleavage and dipping 80° eastward, and bedding commonly dipping 80° westward, a 20° discordancy is produced between bedding and the present position of the orebodies. This transgression of bedding by the sulphide orebodies is a characteristic feature of the Cobar district and has resulted in the repeated emphasis on structural control of the deposits by most previous authors.

Within the district, two main ore types have been recognised and are particularly well developed in the CSA mine. These are (a) siliceous chalcopyrite-pyrrhotite and (b) massive pyrite-sphalerite-galena.

The **siliceous chalcopyrite-pyrrhotite** type contains less than 25 percent (usually 10-12 percent) total sulphides. Copper grades range up to 3.5 percent, and lead and zinc usually amount to less than 0.5 percent each. Adjacent siltstone host rock is usually at least partly silicified and quartz veining is common. Distribution of mineralisation is as disseminations or oriented veinlets of sulphides and quartz in a silicified or chloritized siltstone (Kappelle, 1970; Brooke, 1975, 1976).

The **massive pyrite-sphalerite-galena** ore type tends to contain over 50 percent total sulphide. Pyrite, sphalerite and galena are the most common sulphide minerals; pyrrhotite and chalcopyrite are minor. Lead-zinc ratio ranges between 1:2 and 1:3. The ore is sometimes roughly banded and may grade into massive chalcopyrite-pyrite ore (Kappelle, 1970; Brooke, 1975, 1976). Kappelle (1970, p. 83) in referring to a portion of the western orebody at the CSA mine, reports that the chert there appears to be of chemical origin. Furthermore, he states '. . . mineralization in this chert is most distinctive in that it is conformably controlled by bed-

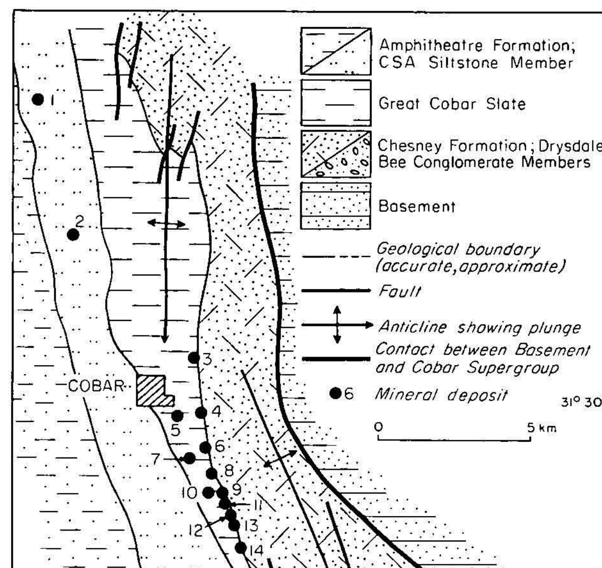

Figure 4. Geological map, Cobar area, NSW. Simplified slightly from Brooke (1975).
1. CSA; 2. Spotted Leopard; 3. Tharsis; 4. East Cobar; 5. Great Cobar; 6. Old Fort Bourke; 7. Dopville; 8. New Cobar; 9. Chesney; 10. Gladstone; 11. Burrabingie; 12. Mount Pleasant; 13. Wood Duck; 14. Occidental; 15. Peak; 16. Coronation; 17. Beechworth; 18. Queen Bee.

ding. The sulphides are well differentiated and segregated into monomineralic bands. . . .'

Relatively minor ore types in the district are a massive siliceous chalcopyrite-pyrrhotite type (the '11 CE type' at CSA; Brooke, 1975) and the massive magnetite-pyrrhotite-chalcopyrite type occurring with stilpnomelane and minor quartz (the 'Great Cobar type'; Brooke, 1976).

The Elura deposit, although not yet described in the literature, would, from personal observation by the writer, be considered to be of the massive pyrite-sphalerite-galena type with minor pyrrhotite.

Geological features of exhalite deposits

The concept of an exhalative origin for many deposits is now accepted by a great number of geologists and a full review of its development history is beyond the scope of this paper. Briefly, Oftedahl (1958) popularised, in English, the exhalative sedimentary theory of ore formation previously expressed by a number of European authors (e.g. Schneiderhohn, 1944; Ramdohr, 1953) in which base-metal accumulations were thought to have been produced by sea-floor fumaroles of volcanic origin. Stanton (1959) strongly supported a volcanic-exhalative origin for ores in the Bathurst area of northern New Brunswick, Canada and later (1960) expanded this concept for all orebodies of the 'conformable pyrite' type.

Geological characteristics of exhalite deposits in volcanic rocks have been described, for example, by Sangster (1972) for certain Canadian Precambrian deposits, by Vokes (1976) for Norwegian Lower Palaeozoic ores, and by Sato (1974) for the Miocene of Japan. Although the exhalative theory has been most widely used for deposits in volcanic terrain, under certain circumstances it can also be advocated in sedimentary terrains. For example Sangster (1972) pointed out that the Sullivan deposit in Canada possessed many of the characteristic features of a volcanic exhalative deposit, yet it occurs in siltstone and shales containing no evidence of volcanism. Whether hosted in sedimentary or volcanic rocks, however, the features which characterise an exhalative deposit are similar in both environments.

Briefly, exhalative deposits consist of two main portions, a sulphide-rich bedded upper portion conformable with bedding in the host rock, and a lower, relatively sulphide-poor discordant zone of veins and/or disseminations enclosed in a broad envelope of hydrothermally altered country rock. The alteration may commonly be any or all of the following: silicification, sericitisation, chloritisation, and carbonatisation. In the author's experience, silicification is the common form of alteration in exhalative deposits enclosed in sedimentary rocks.

The discordant lower zone is invariably poor in zinc and lead relative to the upper portions and, in many deposits, may contain only copper to the virtual exclusion of lead and zinc. This copper zone with its alteration envelope is always stratigraphically below the massive, banded upper portion and is considered to represent the position of the channelway through which metalliferous solutions passed to debouch onto the sea-floor and thereby deposit the conformable massive banded ore at the sediment-water interface.

Between the massive banded lead-zinc ore and the disseminated or vein copper ore, some deposits contain a zone of massive chalcopyrite-pyrite (± pyrrhotite) ore with high copper grades.

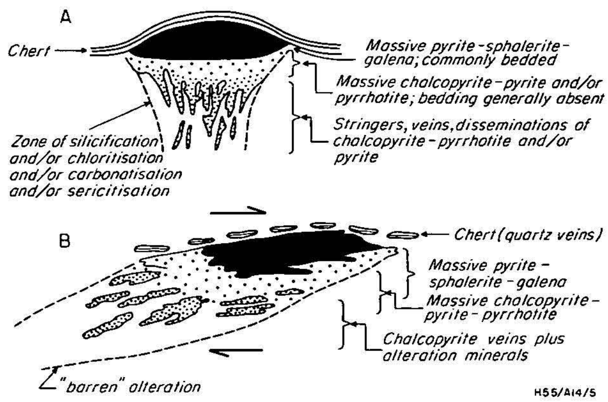

Figure 5A. Schematic diagram illustrating main geological features and ore types of an undeformed exhalative deposit. Not all ore types are present in every deposit. B. Schematic diagram to show effect of transportation of exhalative deposit into plane of schistosity or cleavage. Both diagrams slightly modified from Sangster (1972).

Iron-rich minerals (other than sulphides) such as magnetite, various silicates, and carbonates, may accompany the massive portion of the ores and, in some instances, pass out laterally beyond the 'ore zone' along the favourable horizon. The massive bedded ore commonly contains chert as a gangue mineral; this may occur as a conformable layer of sulphide-free chert immediately above the massive lens.

A schematic cross-section of a typical exhalative deposit is shown in Figure 5A, which illustrates the morphology of an undeformed deposit in terms of the two main ore-types and associated features. Deformation of an exhalative deposit, particularly development of shearing and/or cleavage, will tend to transpose both ore types into the schistosity/cleavage plane (Fig. 5B). Most important, however, is to note that the chalcopyrite-alteration zone, originally markedly discordant to the stratiform pyrite-spaherite-galena portion, is transposed into near-concordancy with it. Individual chalcopyrite lenses will appear in the schistosity/cleavage planes but the zone as a whole will still be discordant to the bedded portion of the orebody, albeit at a small angle. Transposition, however, will not change the relative positions of the bedded versus disseminated ore types; the latter will still be in the stratigraphic footwall.

Application of the exhalative model to deposits of the Cobar district

In presenting the evidence for, and the consequences of, the exhalative origin for Cobar district deposits, data will be presented on both the deposit-scale and the regional-scale.

Deposit-scale evidence

By now the reader will doubtless be aware of the marked similarity between the two main 'ore types' in the Cobar district previously outlined (the siliceous chalcopyrite-pyrrotite type and the massive pyrite-sphalerite-galena type) and the two main ore types of typical exhalative deposits. A critical point in developing the exhalative concept for the Cobar district is to examine the positions of these two main ore types relative to the Cobar district stratigraphy. This can be done using available data from two of the larger depo-

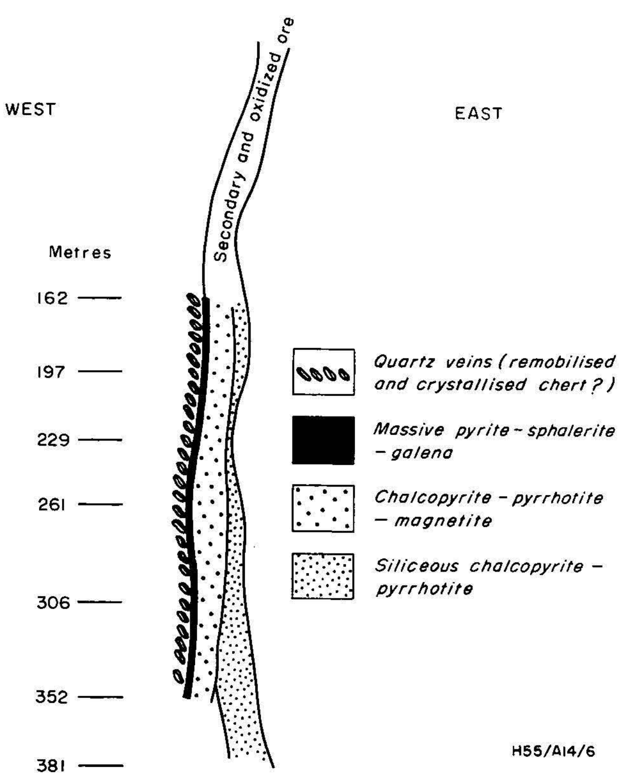

Figure 6. Vertical section, Central Lode, Great Cobar deposit. Distribution of ore types. Simplified slightly from Andrews (1911). Compare with Fig. 4(B).

sits in the area, namely the CSA and the Great Cobar deposits.

The ore distribution of the so-called 'Eastern System' in the 275 m level in the CSA mine is shown in Figure 7A. Note that the massive lead-zinc ore is on the western side of the main zone of siliceous copper ore, indicating a westward-facing sequence in accord with the generally west-facing directions noted for the Cobar Supergroup as a whole. Furthermore, another, smaller lead-zinc lens, slightly west of the main lens, is also on the westward side of an equally small siliceous copper ore zone. Again this indicates a west-facing sequence and, in fact, two 'favourable horizons' are suggested by these two lenses.

The position of the siliceous copper ore relative to the massive lead-zinc ore is in accord with an exhalative origin. The juxtaposition of two ore horizons in the manner shown is a consequence of this interpretation and represents a 'stacked' orebody, a feature not unknown in exhalative deposits.

In the Great Cobar deposit, Thompson (1953, p. 875), in referring to the Central Lens, describes the following ore types from east to west:

(a) silicified slate passing westward into siliceous copper ore;
(b) chacopyrite-pyrrhotite-magnetite ore;
(c) pyritic lead-zinc with a 'well-defined' (i.e. sharp) contact with unaltered slate.

The sequence is shown in Figure 6 from Andrews (1911) and is well documented in detail in a general section across the Central Lode (Andrews, 1911, fig. 2). An analysis of the lead-zinc portion of the lens is reported by Andrews (1911, p. 114) to be: Pb 18%, Zn 19.78%, Fe 14%, and S 23.11%; truly a massive pyritic lead-zinc deposit! The eastern fringe of the lens, in contrast, is described as a network of chalcopyrite and quartz veinlets in silicified slate. In Figure 6, the reader will note how the siliceous chalcopyrite-pyrrhotite 'tail' is discordant to the massive pyrite-sphalerite-galena portion of the deposit. This is a characteristic feature of a sheared or transposed exhalative deposit (compare with Fig. 5B). The disposition of the three ore types demonstrates a westward-facing exhalative deposit.

In addition to these two main deposits in the Cobar area, similar features have been noted in other smaller deposits in the region. For example, at the Chesney deposit, Rayner (1969, p. 62) noted a prevalence of lead and zinc (relative to copper) in the structural footwall ore. In the eastward dipping orebodies, the structural footwall would again be the western side indicating a westward-facing exhalative deposit, a situation identical with the CSA and Great Cobar deposits.

In the New Occidental deposit, Rayner (1960, p. 62) similarly notes a zoning in the deposit with lead-zinc ore occurring on the footwall (i.e. western) side of the lode. This would again be consistent with a west-facing exhalative deposit.

In the Queen Bee deposit, Thompson (1953, p. 886) describes a lead-zinc lode (average assay: 3.3% Zn, 1.9% Pb, 0.4% Cu) on the west side of the main siliceous copper lode (Southern Lode; average grade = 1.3 %Cu).

Similarly, in the Big Lode of the Peak Deposit, Rayner (1958, p. 60) reported assays from a 225-foot drill intersection (from 595'-819') from hanging wall to footwall of the west-dipping body. Ratios of Pb+Zn/Pb+Zn+Cu in the western, central, and eastern portions were 0.88, 0.90 and 0.55 respectively, indicating once more a lead-zinc-rich western side to the lode relative to the eastern side and confirming a westward-facing deposit.

At the Nymagee deposit, Suppel (1976, p. 31) quotes an unpublished report stating that in the Main Lode there is a 'tendency for zinc-lead rich ore horizons to lie to the west and stratigraphically above those that are copper rich and lead poor'. Suppel states that the Nymagee lodes are 'nearly conformable' with the west-dipping, west-facing host sediments. Thus the lead-zinc rich hanging wall and copper-rich footwall in the Nymagee Main Lode is also in accord with an exhalative origin.

In summary then, deposits with lead-zinc 'tops' and copper-rich 'bottoms' have been demonstrated to occur at four horizons in the Nurri Group (Figs. 3 & 4): at the contact between the Bee Conglomerate Member and Chesney Greywacke (Queen Bee deposit); at the Chesney-Great Cobar Slate contact (New Occidental and Chesney deposits); within the Great Cobar Slate (Great Cobar and Peak deposits); and within the CSA Siltstone Member (CSA deposit). In addition, in the Kenmure Group in the Nymagee area (Fig. 1), which may represent a smaller, subsidiary basin separated from the main Cobar basin by a low basement ridge, the Nymagee deposit contributes a fifth horizon or favourable zone exhibiting the same characteristics and relationship to stratigraphy as deposits in the immediate Cobar area.

In view of the evidence of an exhalative origin for deposits of the Cobar district, the average 20° discordancy between bedding and the deposits deserves comment. Relative to most host rocks, particularly siliceous ones such as sandstones, siltstones, felsic volcanic rocks, sulphide minerals by and large are much more plastic and are readily remobilised into planes of movement or zones of decreased pressure. Concentration of sulphides into the hinge zone of folds, for example, is a common feature as is mobilisation of bedded sulphides into faults, shears, and cleavages. Likewise, of all the silicates, silica is perhaps the most readily mobilised and can likewise be reconcentrated into such features as faults, shears, pressure shadows. Bearing in mind this property of sulphides and quartz, the present discordant position of the Cobar district deposits can be readily explained by transposition of sulphides and silica into cleavage planes, presumably as a result of structural deformation of the area. Because of the present steep westward dip of bedding, transposition of sulphides 20° into eastward-dipping cleavages effectively results in structural overturning of the deposits.

Apparently structural conditions in the Cobar area were such that only the orebodies (i.e. sulphides and silica), and not bedding, were transposed. Readers are referred to figure 6 in Brooke (1975, p. 693) showing a piece of drill core in which originally concordant sphalerite, galena, and chalcopyrite, but not bedding in the host rock, have been transposed into cleavage, resulting in a discordancy of 15°-20° between host rock bedding and sulphides. Brooke (1975, p. 694) notes that 'bedded sulphides could have been remobilized along the cleavage during shearing. Recrystallization and deformation textures have been observed in the sulphides, and it is almost certain that remobilization has occurred.' The presence, on the scale of drill core, of sulphides parallel to bedding in the siltstone, as well as discordant to it in the cleavage, is exactly analogous to that described by Brooke (1976) on a deposit-scale (see discussion under Mineral Deposits, p. 18).

Thus it can be demonstrated that, even if they have been transposed into cleavage, the deposits show the same sense of chemical zoning throughout the district in seven deposits at five stratigraphic horizons. The consistency of the facing directions as indicated by this zoning, as well as the internal geological features of the deposits, is in accord with an origin by exhalative processes and later transposition into cleavage.

Regional-scale evidence

Several authors have pointed out the close association of deposits in the Cobar area with iron-enriched sediments. For example, Kappelle (1970, p. 83) states that in the Nurri Group 'disseminated sulphides, mainly pyrite and minor iron-free sphalerite occur throughout'. Brooke (1975, p. 687) notes that 'several linear zones of higher magnetic intensity are roughly parallel to the stratigraphy and most of the orebodies occur within these. The cause of the trends is probably fine-grained pyrrhotite flakes in the sediments, but anomalies at Mopone . . ., Great Cobar, and Dapville are caused by magnetite.' In the CSA mine area, Thompson (1953, p. 887) emphasises that a strong magnetic anomaly was obtained over the concealed orebody and that 'a chain of weak magnetic anomalies persists over a distance of 6 miles southeast of the CSA. . . .' One such anomaly at the old Spotted Leopard Mine was drilled and found to be caused by 'pyrrhotite, occurring as sparse thin flecks in slaty claystone with associated feeble pyrite mineralization'.

The regional extent of the pyrite-pyrrhotite-magnetite disseminated in Nurri Group sediments, together with the occurrence of the orebodies within these iron-enriched sediments, is analogous to the occurrence of the Sullivan Mine, Canada, in pyrrhotiferous siltstones and shales—for which an exhalative origin has also been proposed (Sangster, 1972).

Other iron-rich minerals such as stilpnomelane (an iron-silicate) have been identified from the Great Cobar, Dapville, New Cobar, Chesney, New Occidental, and Nymagee deposits (Rayner, 1969, p. 77). Siderite occurs in the Great Cobar, New Cobar, Dapville, Chesney, Gladstone, and New Occidental deposits (Rayner, 1969, p. 83). Although Rayner does not describe the abundance or the location of these minerals in the deposits, the presence of these iron-rich minerals may be significant in terms of an exhalative origin.

The tendency for the deposits to occur in the finer grained sediments of the Nurri Group has been emphasised by Brooke (1975, p. 688; 1976, p. 28). Presumably he was referring to clastic sediments (i.e. the shales) but the statement could equally well apply to the bedded chert referred to earlier in relation to the Western System of lodes at the CSA mine. In this context, chert would be regarded as a fine-grained chemical sediment. In a sedimentary environment where 'competition' exists between clastic sedimentation and chemical sedimentation, chemical sedimentation will tend to dominate under conditions of reduced clastic input. Presumably the fine-grained clastic sediments were deposited at a lower sedimentation rate than the coarser grained greywackes, sandstones, and conglomerates, thereby permitting the accumulation of chemical sediments such as chert and layered lead-zinc sulphides. The situation is analogous to exhalative deposits in a volcanic environment where the massive sulphide-chert deposits formed during periods of reduced volcanism.

Discussion

Recognition and documentation of evidence for an exhalative origin for deposits of the Cobar district has many implications. One of these obviously is to de-emphasise the traditional concept of structural control for the orebodies in favour of a stratigraphic control.

The nature of mineralisation and compositional zoning in deposits at five stratigraphic horizons in the Cobar-Nymagee area has been shown to be consistent with features of exhalative deposits. The deposits would therefore be compared with other exhalative deposits in non-volcanic terrain such as the Sullivan in Canada and the Rammelsberg in Germany. The original exhalative nature of the Cobar deposits is still evident and, moreover, can be used to define stratigraphy and structure in the Cobar-Nymagee area.

The possible existence of a set of growth faults (which could have provided the channelways for the rising exhalative solutions) contemporaneous with Cobar Supergroup sedimentation, the presence of contemporaneous volcanism (some of it subaqueous and containing possible volcanogenic deposits such as the May Day deposit southwest of Nymagee; Gilligan, 1974, p. 167), and recognition of the original syn-sedimentary nature of mineral deposition, combine to enhance the potential of the Cobar-Nymagee area for Rammelsberg-Mount Isa type of stratiform lead-zinc deposits. The recently discovered Elura deposit is ample evidence of this potential.

Occurrence of base-metal deposits at the four stratigraphic positions in the Nurri Group, referred to previously, need not imply four separate mineralisation events. If the Cobar Supergroup sediments are, in fact, time-transgressive as suggested in Figure 3, then

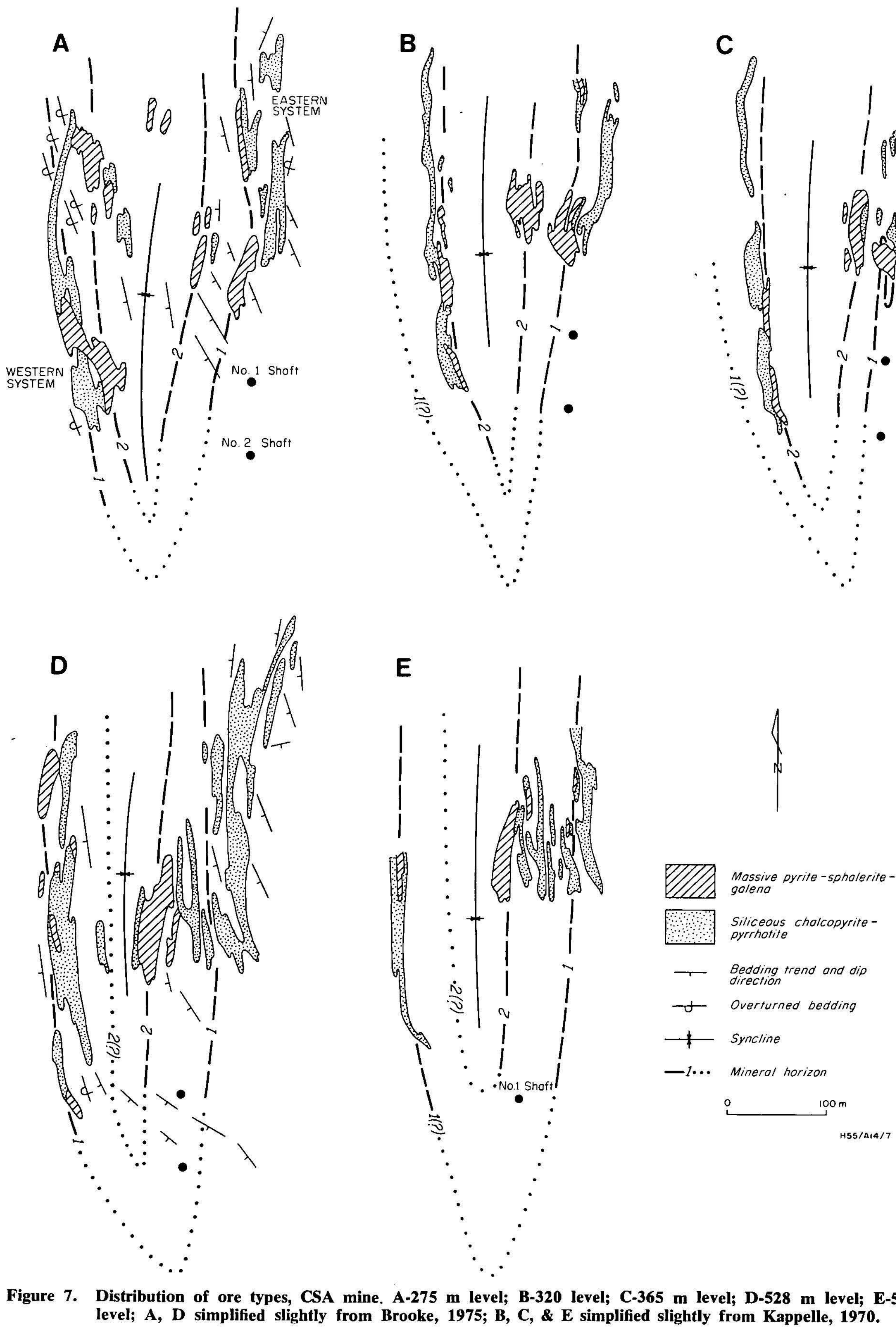

Figure 7. Distribution of ore types, CSA mine. A-275 m level; B-320 level; C-365 m level; D-528 m level; E-550 m level; A, D simplified slightly from Brooke, 1975; B, C, & E simplified slightly from Kappelle, 1970.

mineralisation at all five stratigraphic positions, when the Nymagee deposit in the Kenmure Group is included, could be essentially coeval.

On a more local scale, such as within a single mine, recognition of the sedimentary control of mineral deposition can be used to advantage in searching for new lenses. An example would be the CSA mine where, using published mine plans together with the exhalative concept, it can be demonstrated that the so-called 'Eastern and Western' Systems are, in fact, the two limbs of a syncline. The western limb is, at present, east-facing so, in order to revert to the generally regional west-facing situation, the possibility of the same mineralised horizon being repeated west of the present Western System can be contemplated.

This concept is illustrated in the series of mine level plans shown in Figure 7A-E. In these diagrams, lead-zinc lenses at two stratigraphic horizons (labelled 1 and 2 in the diagrams) with their siliceous copper zones in the footwall can be recognised in a normal west-facing position on the east side of the level plans. The 'Western System' bodies, also on two stratigraphic horizons, are, by contrast, east-facing as shown by zoning of the lenses on the various mine levels. Kappelle's diagram (1970, p. 93) is particularly convincing in showing the east-facing nature in this limb of the fold. His diagram of the sulphide mineralisation in the western orebody on the 365 m level of the CSA mine is reproduced in Figure 8. Note the westward succession from pyrite to pyrite-sphalerite-galena to pyrite-chalcopyrite-sphalerite to chalcopyrite-pyrrhotite types of mineralisation.

Although neither Kappelle (1970) nor Brooke (1975) mention the presence of an isoclinal fold in the CSA mine and, in fact, state that all bedding, where seen, is west-facing, both authors note (Kappelle, 1970, p. 82; Brooke, 1975, p. 689) that, in the vicinity of the ore zones, wallrock alteration has obliterated most of the rock distinguishing features, including, presumably, criteria for top determination in the sediments. Both authors also remark on the absence of distinctive marker beds in the mine which could be used to define large-scale structures. The present author suggests that both these features (facing direction and marker beds) are inherent in the ore zones themselves.

If the facing directions, and hence the syncline, illustrated in Figure 7A-E are accepted then, to comply with the general west-facing direction of the Cobar district, it follows that the western limb must reverse direction somewhere west of the present 'Western System' of lodes, and hence repeat the two favourable horizons now present on both the eastern and western limbs of the CSA fold.

Any evaluation of the exhalative concept as applied to the Cobar area must take into account the compositional and mineralogical zoning described independently by several authors over the past sixty years. The asymmetric nature of this zoning, with the relatively zinc-lead rich portion consistently occurring on the western side of the lodes, if not accepted as evidence of an exhalative origin, must be satisfactorily explained by those who oppose the syn-sedimentary concept.

Similarly, the author is fully aware that the evidence presented for an isoclinal fold in the CSA mine is anything but conclusive, based as it is on data collected and presented by other authors for other purposes. Nevertheless, the reason for the reversal in zoning as expressed in the Western System of lodes in the CSA mine demands explanation and the syncline concept is offered herein as one possibility.

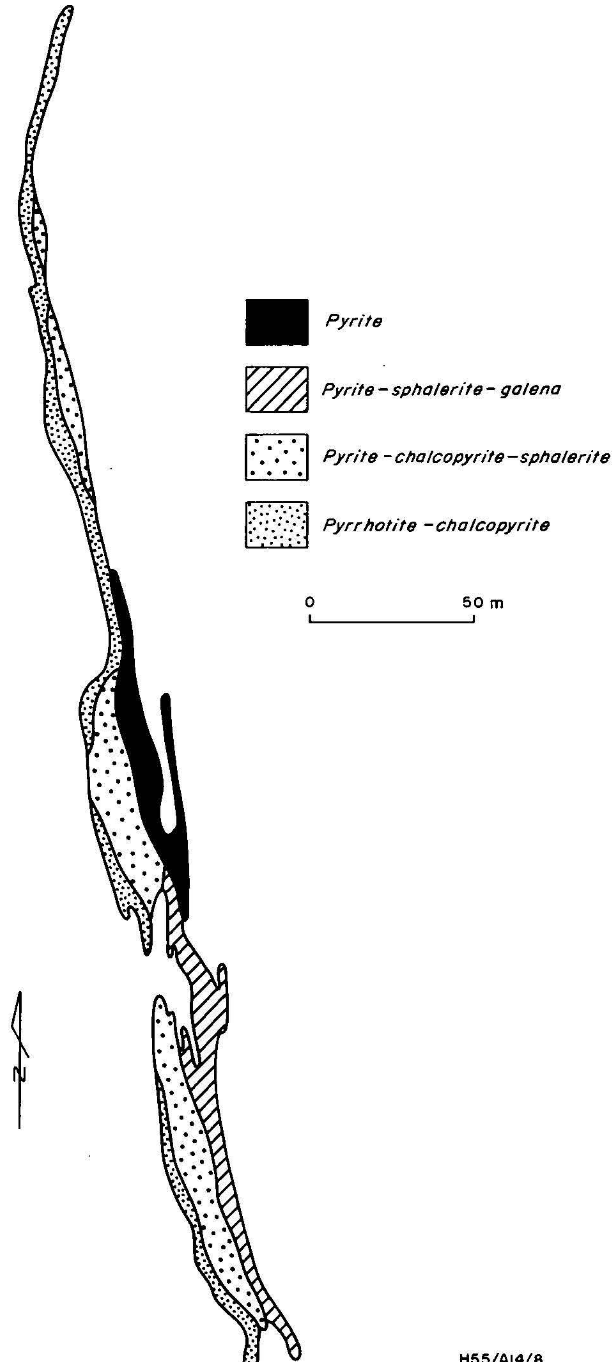

Figure 8. Sulphide mineralisation, western orebody, 365 m level, CSA mine. Note the remarkable mineralogical zoning indicating an eastward-facing exhalative deposit. Slightly simplified from Kappelle (1970).

Acknowledgements

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To all these people, the author is grateful for their assistance. This does not, however, imply that they necessarily agree with either the method or the conclusions presented. For this, the author bears full responsibility.

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Water levels, balance, and chemistry of Lake George, New South Wales

G. Jacobson & A. W. Schuett

The marked water-level fluctuations of Lake George, a closed lake in southern New South Wales, have long been a subject of speculation. Monitoring over a 20-year period (1958-77) shows that the fluctuations are a response to seasonal and long-term variations in rainfall, evaporation and inflow of streams. An approximate water balance for the lake has been computed and shows marked seasonal characteristics; increases in water volume between May and October correspond with high inflows and low evaporation, while decreases in water volume between November and April correspond with low inflows and high evaporation. The long-term fluctuations reflect climatic variability. Salinity of the lake water, which is a sodium chloride type, varies inversely with water volume. A substantial net loss in salt was observed during a recessive phase of the lake in 1973.

Introduction

Lake George is a closed lake in a basin of internal drainage covering 932 km^2 in the Southern Tablelands of New South Wales (Fig. 1).

The lake was 'a splendid sheet of water' when first discovered by European colonists in 1820, but by 1839 'was dry enough to drive a team across the middle (historical sources quoted by Russell, 1887). The considerable fluctuations in the water level of Lake George have created a good deal of interest ever since. Thus, Griffith Taylor wrote in 1907—

'The lake is now portioned into grazing leases, and fences run nearly across the bed. The local sheep-breeders for the most part much prefer the lake dry, since many extra sheep can be carried on their runs . . . At the same time the neglected boathouses,

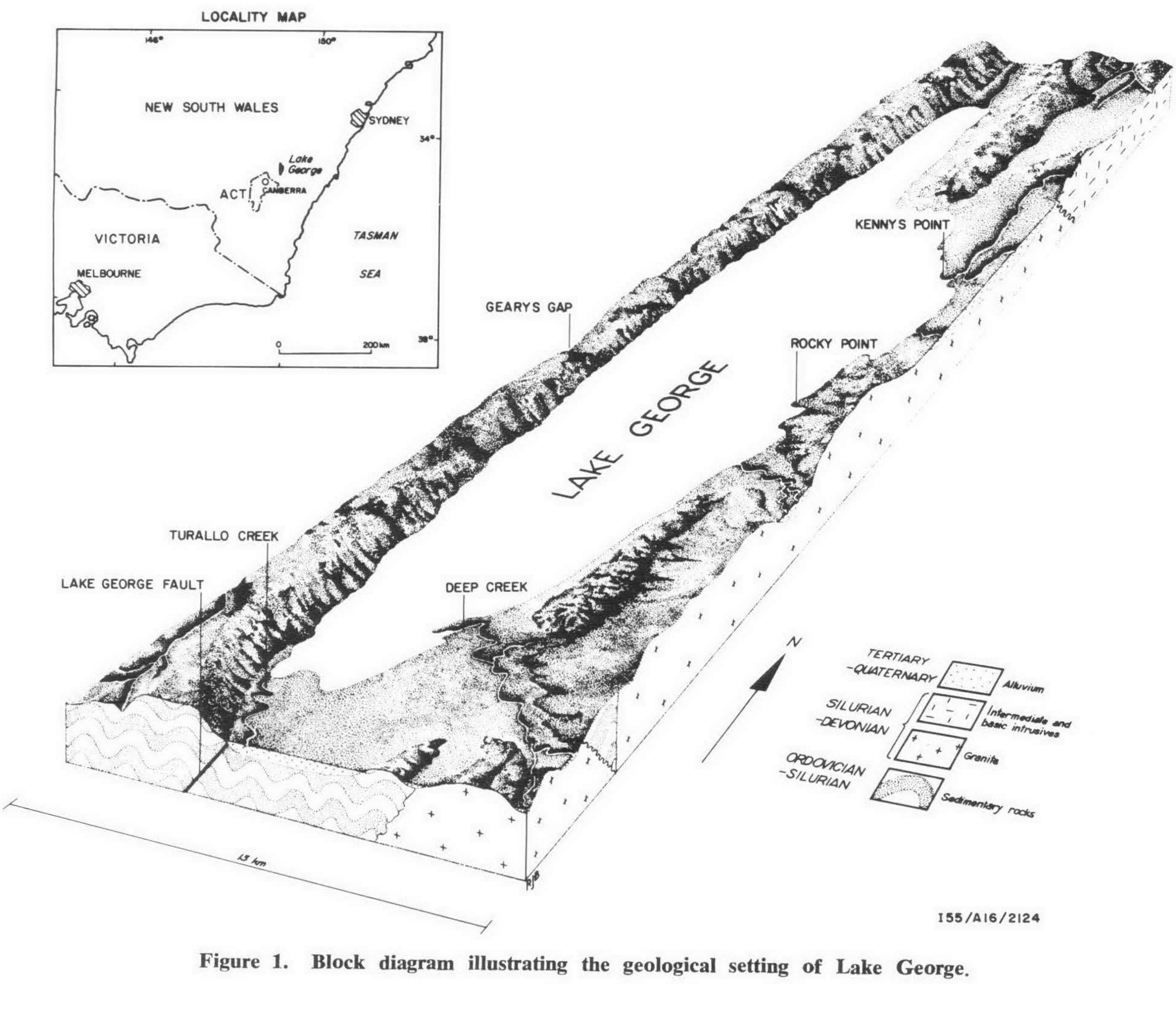

Figure 1. **Block diagram illustrating the geological setting of Lake George.**

Figure 2. Lake George looking southeast from Geary's Gap, March 1978.

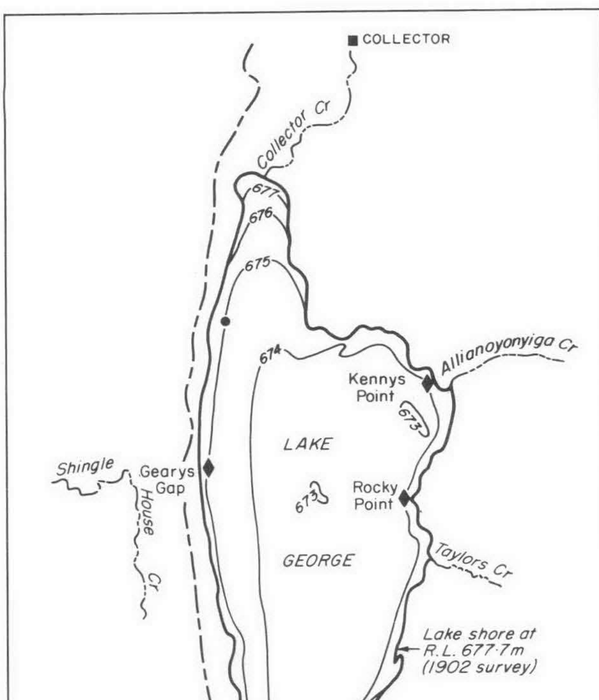

Figure 3. Lake George: contours on the lake bed.

jetties, and decaying boats and launches which are to be seen near Bungendore, recall the good old times when the lake teemed with Murray cod . . .'

A substantial body of myth has grown up to explain the appearing and disappearing waters of Lake George. Popular theories have included the draining of lake water through a fissure and its subsequent refilling from an underground spring! Recently, because of the lake's proximity to Canberra, possible recreational and water storage uses have been mooted for it, and the study of its fluctuations is thus of economic importance.

The hydrographic record of Lake George extends over 160 years and is one of the longest in Australia. The first official gauge at Lake George was set up in 1885 by H. C. Russell, the Government Astronomer of New South Wales (Russell, 1887). Levels were gauged by the New South Wales Department of Works until the lake dried up in 1930. BMR has gauged Lake George since 1950 (Noakes, 1951) and detailed monitoring for computation of the water balance has been undertaken since 1958 (Burton, 1972; Burton & Wilson, 1973).

This paper summarises the results of BMR monitoring of the lake over the last 20 years.

Geological setting

The geological setting of Lake George is shown in the block diagram (Fig. 1).

The lake is underlain by Tertiary-Quaternary alluvium which overlies deeply weathered Palaeozoic bedrock. Geophysical surveys (Polak & Kevi, 1964), and drilling to the south of the lake, have shown that the alluvium is 45-50 m thick; it consists of interbedded sand, clay and gravel. A drillhole in the lake bed near the north end of the lake was completed in mainly clayey alluvium at a depth of 71 m. The alluvium around the lake is an important source of sand and gravel for Canberra development, and contains aquifers which supply water to the town of Bungendore and the Woodlawn mine.

Geophysical surveys have shown that bedrock beneath the lake is weathered to depths of up to 130 m. The bedrock consists mainly of Ordovician and Silurian sedimentary and granitic rocks (Garretty, 1936). These rocks are separated from less weathered Ordovician rocks to the west by the Lake George

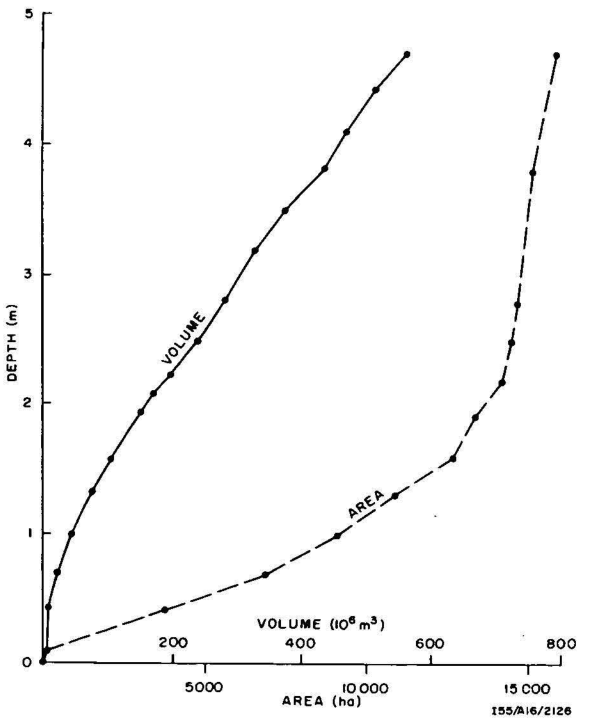

Figure 4. Area and volume of water in Lake George as a function of maximum depth.

Fault, a steeply dipping, meridional, reverse fault (Noakes, 1957). The escarpment on the west side of the lake is probably the result of post-Palaeozoic movement on the Lake George Fault, and the lake probably originates from disruption of a pre-existing drainage system by this movement (Jennings, Noakes & Burton, 1964).

Abandoned shoreline features up to 37 m above the lake bed indicate a history of substantial water-level fluctuations dating back beyond 20 000 years B.P. (Coventry, 1976). Above this height, the lake would have overflowed to the west through a saddle in the escarpment known as Geary's Gap. Pollen studies of lake bed cores up to 9 m deep, by Dr G. Singh of the Australian National University, have extended the history of vegetation and lake conditions back possibly 100 000 years (Bowler & others, 1976; Churchill & others, 1978).

Water-level fluctuations

Monitoring of water levels, temperature and salinity has been carried out monthly from one or more of the three gauging stations, Geary's Trig., Kenny's Point, and Rocky Point (Fig. 3). The lake is subject to fluctuations caused by wind-generated seiches with an amplitude of up to 15 cm and a period of up to 2 hours (Burton, 1972).

The lake bottom is smooth with the lowest point just below 673 m (Fig. 3). Lake-bed surveys in 1902 and 1973 showed only slight changes of up to 15 cm due to sedimentation during this time. The lake area and volume depend on the shape of the basin and are related to maximum depth as shown on the graph in Figure 4.

The results of monitoring of the water-level fluctuations of Lake George between 1958 and 1977 are shown in Figure 5. There are regular seasonal changes in lake level in response to rainfall and evaporation, typically with maxima in October and minima in April. The seasonal fluctuations are superimposed on long-term variations with a period of several years. The highest level reached during the period of measurement was 677.29 m, a maximum depth of 4.41 m, in October 1963; this corresponded to a lake area of 15 400 hectares, and a volume of just over 500 million cubic metres. The lowest level reached was 673.47 m, a depth of 0.59 m in July 1973. This corresponded to a lake area of 4 300 hectares and volume of 7 million cubic metres, and was the culmination of a low stage lasting several years.

The composite lake hydrograph based on historical records from 1818 to 1977 is shown in Figure 6. Depths of more than 7 m were recorded in 1820-23 and 1873-75 (Russell, 1887). Rainfall records are available for Bungendore from the 1880's. From the rainfall variation graph (Fig. 6), it is evident that the periods of 1904-12 and 1930-45—when the lake was dry or nearly dry—were periods of below average rainfall in the catchment.

Water balance

The water balance derived from Lake George is based on the simplified equilibrium equation for closed lakes,

$$I + P\,(A_L) = E\,(A_L)$$

where I is inflow from the catchment, P is precipitation on the lake surface, E is the gross evaporation rate, and A_L is the lake surface area (Langbein, 1961). Variations in inflow, precipitation, and evaporation produce fluctuations in water level with corresponding changes in lake area and volume. This can be represented by the equation

$$\Delta S = I - (E\text{-}P)\,A_L$$

where ΔS is the change in lake volume.

Groundwater seepage into and from Lake George is thought to be only a minor component of the water balance, and has not been taken into account in computations. Piezometric levels in two observation bores about one kilometre south of the lake are generally 6-7 m above lake level, which indicates a net groundwater inflow to the lake.

The approximate water balance for Lake George has been calculated for the past 20 years (Table 1). The data used for computation are precipitation and evaporation which are based on meteorological station data, and the monthly change in lake level obtained at the gauging stations in the lake. Inflow from the catchment is not gauged, but is estimated by difference; any errors in the components of the water balance are thus incorporated into the inflow component.

Precipitation is taken as the average monthly rainfall for two rain gauges in the catchment at Bungendore and Collector. The mean annual precipitation on the lake is estimated as 749 mm; the annual precipitation has fluctuated between 392 mm (1967) and 1119 mm (1974). Precipitation is distributed fairly evenly throughout the year, with a maximum in October. The total precipitation on the lake is estimated as about 100 million m^3 annually.

The evaporation component is based on monthly measurements at the Canberra meteorological station, 30 km to the southwest of Lake George. Until February 1967 Canberra evaporation was measured with an Australian Sunken Tank; Burton (1972) established that these measurements are very close to the actual lake evaporation. After February 1967, Canberra

	Monthly rainfall (mm)	*Change in lake level (mm)*	*Change in lake volume (million m³)*	*Volume of rain on lake (million m³)*	*Evaporation from lake*		*Inflow to lake (million m³)*
					Volume (million m³)	*Depth (mm)*	
January	69	— 83	—10.59	9.13	23.56	177	3.83
February	66	— 42	— 8.21	8.40	18.47	138	1.85
March	47	— 70	— 9.21	6.22	15.98	121	0.56
April	56	— 24	— 2.46	7.00	10.02	77	0.56
May	51	+ 2	+ 0.96	6.61	5.82	45	0.17
June	43	+ 40	+ 4.51	5.73	3.76	28	2.54
July	56	+ 79	+ 9.32	7.63	4.20	32	5.89
August	69	+ 85	+11.26	9.02	6.36	47	8.59
September	65	+ 40	+ 9.14	9.12	9.42	70	9.45
October	97	+ 70	+ 7.33	13.42	14.80	109	8.71
November	65	— 30	— 5.97	8.76	21.68	157	6.95
December	65	—121	—13.10	8.84	24.22	183	2.28
Annual Total	749	— 54	— 7.02	99.88	158.29	1184	51.38

Table 1. Approximate mean monthly water balance, 1958-77.

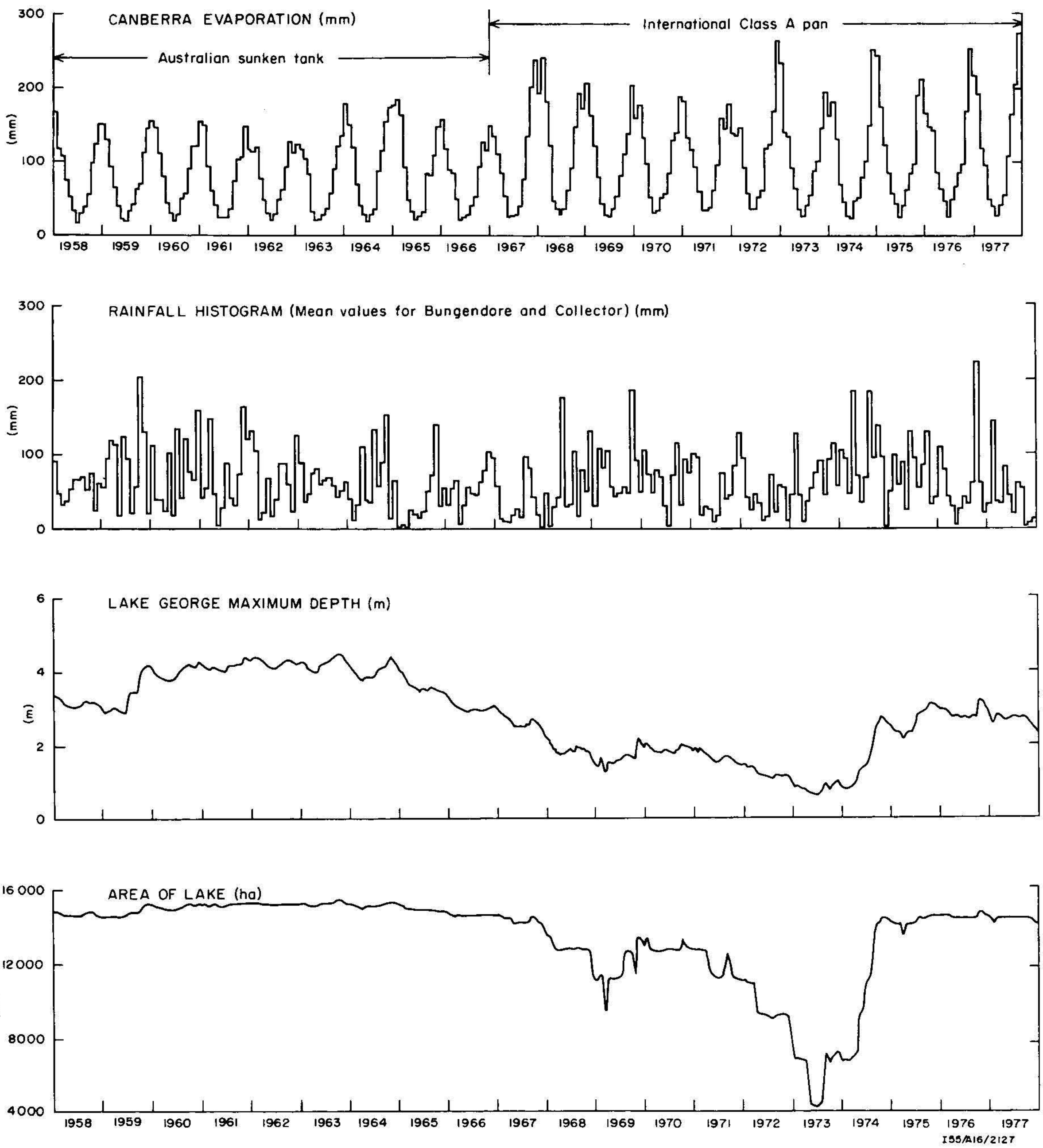

Figure 5. Fluctuations of Lake George 1958-1977.

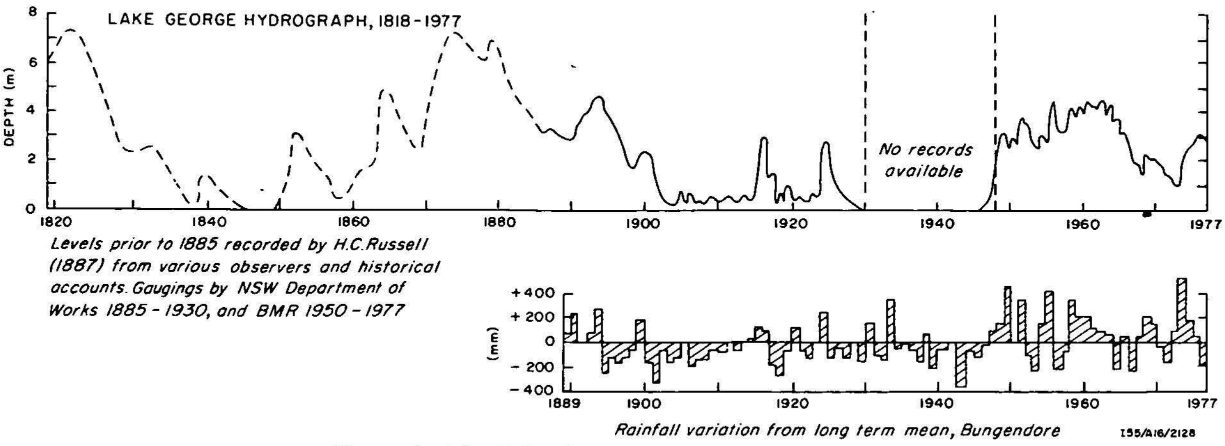

Figure 6. The Lake George Hydrograph 1818-1977.

evaporation was measured with an International Class A pan, and a correction factor of 0.85 has been applied to these pan measurements to estimate the lake evaporation. This value is the mean of experimentally determined annual pan coefficients, relating pan to lake evaporation at two water storages in New South Wales (Hoy & Stephens, 1977). A birdguard was added to the pan in April 1975; to allow for this a correction factor of 0.90 has been applied to more recent pan measurements.

The mean annual evaporation from the lake is estimated as 1184 mm per year and has ranged between 987 mm in 1963 and 1650 mm in 1977. The lake water losses from evaporation are estimated as 158 million m^3 annually, equivalent to the total volume of the lake at a level of 2.00 m.

The mean annual inflow of streams is estimated by subtraction from the water balance as about 51 million m^3. Of the streams that flow into Lake George two, Deep Creek and Collector Creek (Fig. 3), have nearly permanent flows; the other three, Taylors Creek, Turallo Creek, and Allianoyonyiga Creek, flow only after heavy rain.

The aggregate loss in water volume between 1958 and 1977 was about 7 million m^3, and the annual volume change fluctuated between a loss of 86 million m^3 in 1967 and a gain of 211 million m^3 in 1974.

Table 1 illustrates the marked seasonal influence on lake volume: from May to October the lake volume generally increases with the relatively high inflows during a time of low evaporation, whereas from November to April the lake volume decreases, with low inflows during the period of high summer evaporation.

Water chemistry

The relationship of the salinity of Lake George to the volume of lake water is illustrated in Figure 7, which is based on monthly electrical conductivity readings at the BMR gauging stations over the past 20 years. The salinity, in terms of total dissolved solids expressed as milligrams per litre, is related to electrical conductivity measured in microsiemens per centimetre at 25°C, by a factor ranging from 0.53 to 0.84—as shown by numerous chemical analyses of the lake water. For purposes of computation the ratio of total

Location / *Date*	*Lake George Geary's Trig Jan. 1969*	*Lake George Kenny's Point Jan. 1969*	*Lake George Geary's Trig Feb. 1973*	*Lake George Kenny's Point Feb. 1973*	*Lake George Geary's Trig Sept. 1976*	*Lake George Kenny's Point Nov. 1976*	*Lake George Geary's Trig Nov. 1976*	*Deep Creek Nov. 1976*	*Collector Creek Nov. 1976*	*Bore South end Lake George Nov. 1976*
pH	8.9	8.9	8.9	8.9	8.5	8.0	7.8	7.4	7.5	8.0
TDS	9893	7163	44800	38900	2018	1627	1730	205	406	432
EC	16250	11900	54052	45953	3601	3071	3251	337	844	854
Ca	18	22	35	35	53	19	22	14	38	42
Mg	205	152	860	740	42	36	38	15	39	21
Na	3506	2504	15900	13700	670	553	585	37	57	103
K	17	13	47	40	6	5	6	2	1	2
Fe	—	—	0.5	0.35	—	0.20	0.10	0.55	0.15	—
Mn	0.02	0.03	0.17	0.17	—	—	—	—	—	0.13
B	—	—	0.9	1.3	—	0.20	0.20	0.15	—	—
F	1.88	1.54	1.1	1.05	—	0.45	0.50	0.20	0.10	0.65
Cl	4995	3535	23005	20160	977	781	830	40	157	103
SO_4	470	325	2823	2391	114	92	97	43	33	11
HCO_3	1085	930	1565	1515	279	251	263	101	148	302
CO_3	66	33	565	450	11	—	—	—	—	—
PO_4	0.94	1.01	1.09	1.04	—	—	—	—	—	—
SiO_2	4.56	3.14	4.5	1.3	—	3	2	13	5	17
NO_3	0.16	0.10	0.25	0.25	7	8	13	3	8	1
NO_2	—	—	0.03	0.03	—	—	—	—	—	—
Br	23	16	100	90	—	6	7	—	—	—
Al_2O_3	—	—	0.40	0.40	—	0.20	0.15	0.58	0.08	—
Cu	0.02	0.02	0.10	0.05	—	—	—	0.04	—	—
Zn	0.04	0.01	0.07	0.12	—	0.01	0.01	0.01	0.02	0.36
Sr	0.75	0.72	2.70	2.35	—	0.35	0.35	0.10	0.20	0.40

Table 2. Chemical analyses of water samples.

Analyses in milligrams/litre with electrical conductivity (EC) in microsiemens/cm at 25°C.

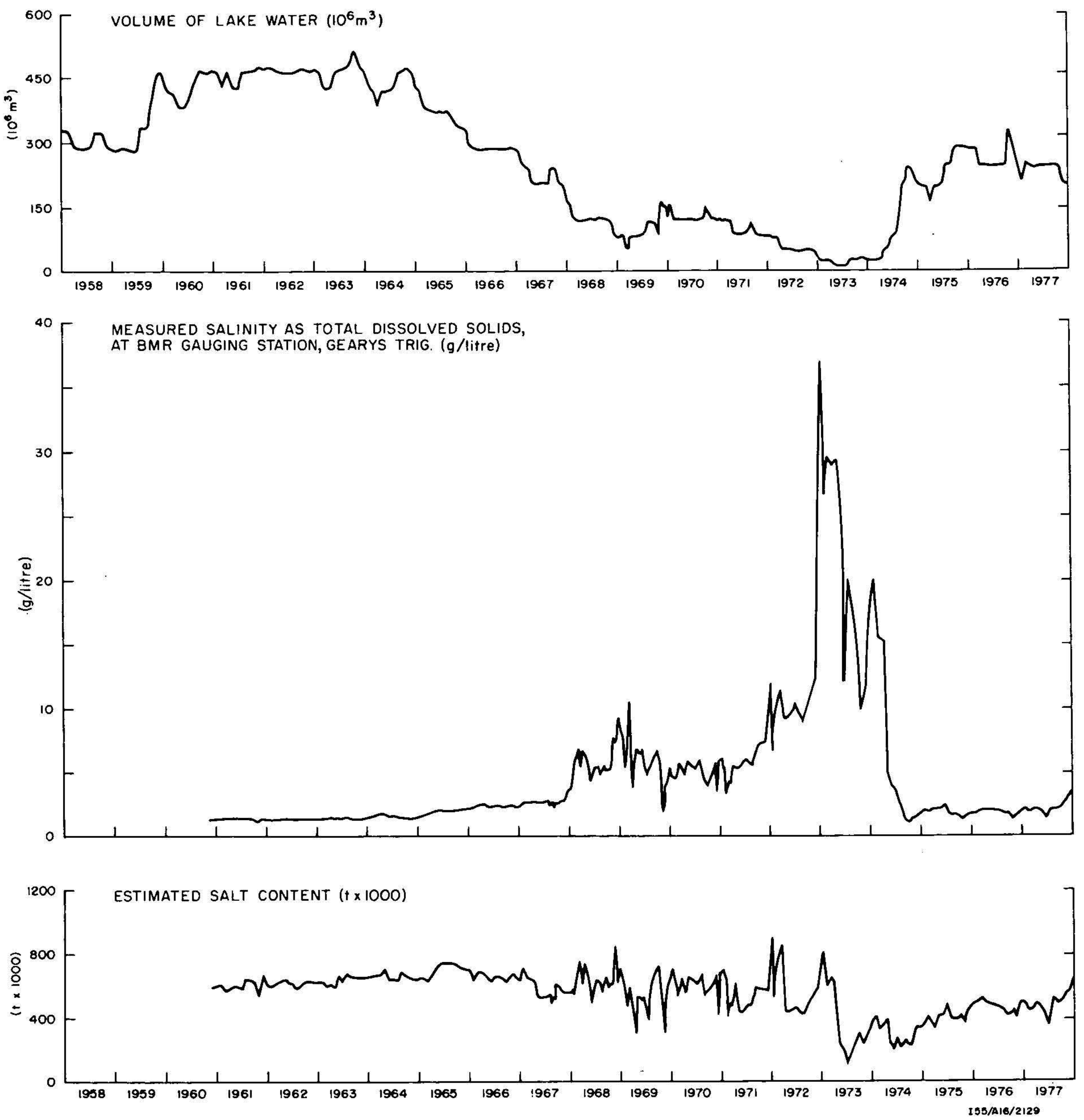

Figure 7. Salinity and salt content of Lake George waters 1958-1977.

dissolved solids to electrical conductivity has been taken as follows.

EC (microsiemens/cm)	TDS/EC
1- 5 000	0.58
5 001-20 000	0.64
20 001-40 000	0.74
40 001-60 000	0.84

Figure 7 shows that the salinity of the lake water increases as the depth and volume decrease, and seasonal fluctuations in salinity are superimposed on long-term trends. During the period for which regular measurements have been made the estimated salinity has ranged from 1183 mg/*l* total dissolved solids in October 1961, to 36 960 mg/*l* in January 1973.

Table 2 shows the results of chemical analyses of lake water at different salinity levels, and of creek water and groundwater. The lake water is generally more saline than creek water and groundwater in the catchments. It is a sodium chloride water, in contrast to the influent creek water which has relatively high concentrations of other ions. The ionic composition of the lake water varies slightly with changes in lake volume; as the lake volume decreases and salinity increases, the concentrations of sodium and chloride increase with respect to other ions.

Traverses across the lake in December 1970 and January 1971 showed an increase in salinity of 6 percent from east to west. Generally fresher water on the eastern side is attributed partly to greater depth and partly to the inflow of fresh runoff from the creeks on the eastern side. Samples recovered from different depths at particular localities showed no appreciable differences in salinity.

The total salt content of the lake (Fig. 7) has been calculated as the product of the weight of water and the estimated average concentration of dissolved solids. Figure 8 shows the variations in salt content of Lake George with changes in lake depth and volume from 1970 to 1977, during which period the lake fell to a

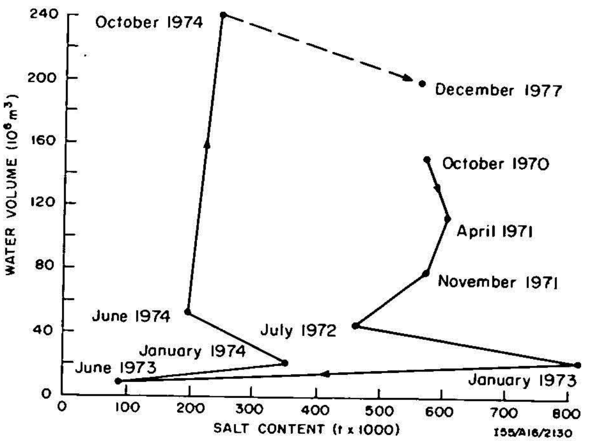

Figure 8. Lake George: relationship of water to volume to salt content, 1970-1977.

low level in July 1973 and then recovered. During the substantial decrease in lake volume from 1970 to 1972 the amount of salt in solution remained approximately constant, although salinity increased markedly during this time. In 1973 an additional small decrease in volume resulted in a marked decrease in the salt content of the lake, from 800 000 to 100 000 tonnes. With the recovery in lake volume during 1974 the amount of salt increased—but not to the same concentration as before the recession began. Subsequently, during the relatively stable phase of the lake from 1975 to 1977 the amount of salt in solution steadily increased from 250 000 to 600 000 tonnes.

The loss of salt during the recessive phase of the lake is attributed to its removal from dry salt flats by winds and dispersal over the surrounding area.

Discussion

The fluctuations in water level of Lake George can be considered on three time scales—daily, seasonal, and over a period of several years. The daily fluctuations are caused by wind-generated seiches. The seasonal fluctuations are due to high inflows and low evaporation raising the water level in the winter months, and low inflows and high evaporation lowering the water level in the summer months. The water-level fluctuations with a period of several years (Fig. 5) are caused by climatic changes, which alter the balance between inflow, evaporation, and precipitation.

Statistical analyses of rainfall data for southeast Australia have shown a general decrease in rainfall from the 1890's and a return to wetter conditions in the 1940's (Gentilli, 1971). These long-term rainfall fluctuations have been ascribed to changes in atmospheric circulation by Kraus (1955, 1962), who postulated a shift in the westerly wind belt towards the equator as the cause of the rainfall decrease in the 1890's. The long-term rainfall fluctuations correlate with two major events at Lake George in historical times—the drying up of the lake in the 1890's, and its refilling in the late 1940's (Fig. 5). The lake hydrograph is therefore an important indicator of gross climatic change.

Geomorphological evidence of high lake levels has enabled extrapolation of the Lake George hydrograph back into Quaternary time to form the basis of palaeoclimatic reconstructions. Thus, Galloway (1965) was able to deduce low evaporation during the last glacial stage and postulate a cooler drier climate than the present. Coventry (1976) has confirmed this, and dated the sequence of abandoned shorelines to plot a history of climatic fluctuations for the last 40 000 years.

Salinity of the lake water varies inversely with water volume, and seasonal fluctuations in salinity are superimposed on long-term trends. The total amount of salt in solution varies in a more complex manner with a cycle of accumulation, removal, and further accumulation (Fig. 8) which conforms closely to the idealised cycle for variations in the salt content of a closed lake postulated by Langbein (1961). The salt in Lake George is thought to be mainly cyclic salt (Burton, 1972), brought in by coastal rainfall. The predominantly sodium chloride composition of the lake water is similar to that of most other southern Australian lakes for which data is available (Williams, 1966).

Any disturbance of the natural hydrological system will require careful analysis and planning to ensure acceptable water level and salinity fluctuations.

Acknowledgements

The late G. M. Burton initiated this project, and developed the computer program used to estimate the water balance. Meteorological data was supplied by the Bureau of Meteorology, Canberra. Chemical analyses of water samples were done by the Australian Mineral Development Laboratories, Adelaide. The lake bed was surveyed by the New South Wales Public Works Department in 1902, and the Australian Survey Office in 1973. The photograph (Fig. 1) was taken by J. E. Zawartko, and Figures 2-8 were drawn by R. Baldwin and G. Clarke. We thank Prof. J. N. Jennings, W. H. Williamson and E. G. Wilson for reviewing the manuscript.

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The stratigraphic sequence of old land surfaces in northern Queensland

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In northern Queensland there are several old land surfaces which are related to previous periods of erosional and depositional activity. The surfaces are of two main types—buried and exhumed unconformity surfaces formed during periods of active erosion and deposition; and terminal planation surfaces of both erosional and depositional origin which formed as stability returned at the end of each cycle of erosional activity. These terminal surfaces have generally been deeply weathered or duricrusted. The oldest landforms are exhumed unconformity surfaces of Cambrian and Mesozoic age which formed the base of the Georgina and Carpentaria Basins respectively. Three sets of unconformity and terminal surfaces are related to cycles of activity within the Cainozoic Karumba Basin. They have been correlated with surfaces elsewhere in Queensland and the Northern Territory. There are also subsidiary depositional surfaces of the current cycle, and a suite of arid landforms which has previously been treated as a 'surface'.

Introduction

This paper defines and discusses the nature and stratigraphy of old land-surfaces in the Cainozoic Karumba Basin, and in the Burdekin Uplands to the southeast (Fig. 1). It reports one aspect of the study of the Mesozoic Carpentaria and Cainozoic Karumba Basins by a combined Bureau of Mineral Resources and Geological Survey of Queensland team, which worked in the area between 1969 and 1974. The geology of the Karumba Basin is discussed by Doutch (1976) and Smart & others (in prep.). This study of the land surfaces builds on the earlier work of King (1949), Twidale (1956, 1966), Galloway & others (1970), and Wyatt & others (1970, 1971).

The distribution of the old land surfaces is shown in Figure 1, based on plate 2 of Smart & others (in prep.). The southeastern part of this figure is outside the area studied by the field party and the distribution of surfaces in this area is based on an interpretation of air photos and previously published geological maps and reports, together with a brief field reconnaissance. Correlation of surfaces between this area and the Karumba Basin is only tentative, and a separate nomenclature is therefore maintained. The Cairns hinterland has not been included in this study. Its geomorphic evolution is interpreted by de Keyser & Lucas (1968).

In referring to physiographic regions, the nomenclature of Smart & others (in prep.) is followed. This incorporates many of the units originally named by Twidale (1966).

Terminology and concepts

Some of the old land surfaces within the area have been named by earlier workers; the nomenclature is further formalised here. In naming the surfaces the use of age terms has been avoided, as this can lead to problems if the chronology of the surfaces is later revised. Although a relative sequence of events is fairly well established in the region, the absolute ages of the surfaces are still uncertain.

The old land surfaces within the study area can be classified into two major and several subsidiary types. The two major types are genetically related to erosion cycles within the region. The concept of these cycles is discussed in Smart & others (in prep.). In general each cycle was started by uplift at the basin margins, or by some other disturbance, and its initial phase was one of active erosion at the margins and deposition in the adjoining depressions. In time, erosion, and therefore deposition, became less active as the source areas were worn down and the depressions filled as the cycle entered its final phase. Eventually, a terminal surface of low relief appeared and deep weathering of both its erosional and depositional parts became the dominant process. The development of the Mesozoic Carpentaria Basin was one such cycle, and three cycles occurred in the evolution of the Cainozoic Karumba Basin (Smart & others, in prep.).

The two main types of old land surface recognised in the area are the unconformity surfaces which formed and were buried during the initial active phase of a cycle; and the terminal planation surfaces which appear during the final stable phase of each cycle.

The **unconformity surfaces** are generally erosional surfaces, buried beneath the sediments of an expanding or migrating depositional area—though in the central part of the basin, where deposition commenced early in the cycle, downwarped remnants of earlier surfaces may have been buried and preserved with little erosion. The buried unconformity surfaces are generally diachronous, as erosional development was continuing in some areas at the same time as depositional preservation elsewhere. The oldest parts are generally near the basin centre where deposition first commenced, and the youngest parts near the basin margin, where burial only occurred late in the cycle. This concept is illustrated in Figure 2, in which the age of the surfaces is shown on the vertical scale and geographic position on the horizontal scale.

Generally unconformity surfaces remain buried, and their shape can only be interpreted in general terms from borehole data or geophysics. For example, Grimes & Doutch (1978, fig. 5) have mapped broad valleys and ridges of the Sub-Claraville Surface in this manner. Exposure by later erosion will generally destroy the surface, leaving only the unconformity trace between the older and younger rocks. However, if the rock below the unconformity is more resistant to erosion than the sediments above it, then the surface may be exhumed with little modification.

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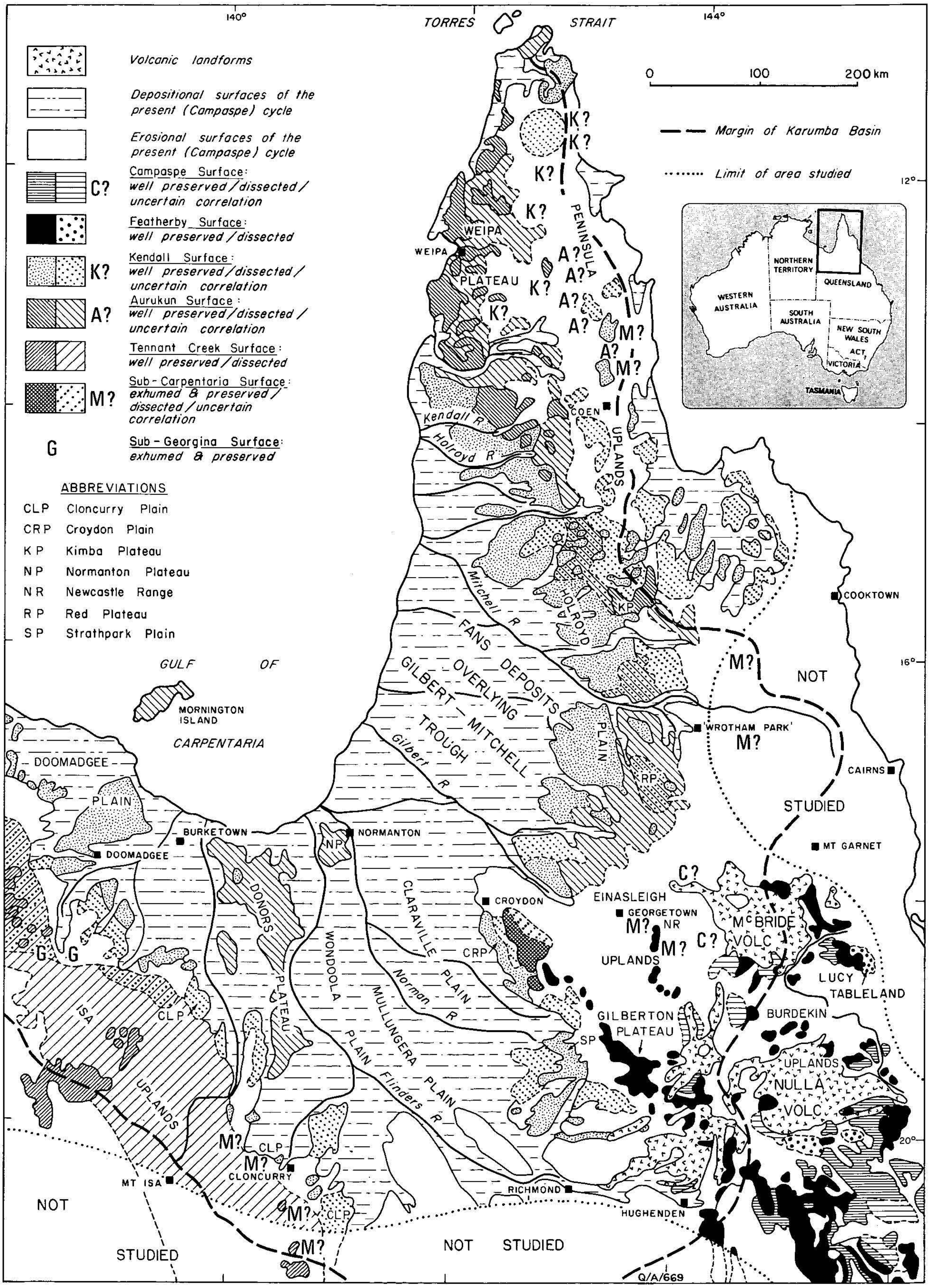


Figure 1. Present distribution of old land surfaces in northern Queensland.

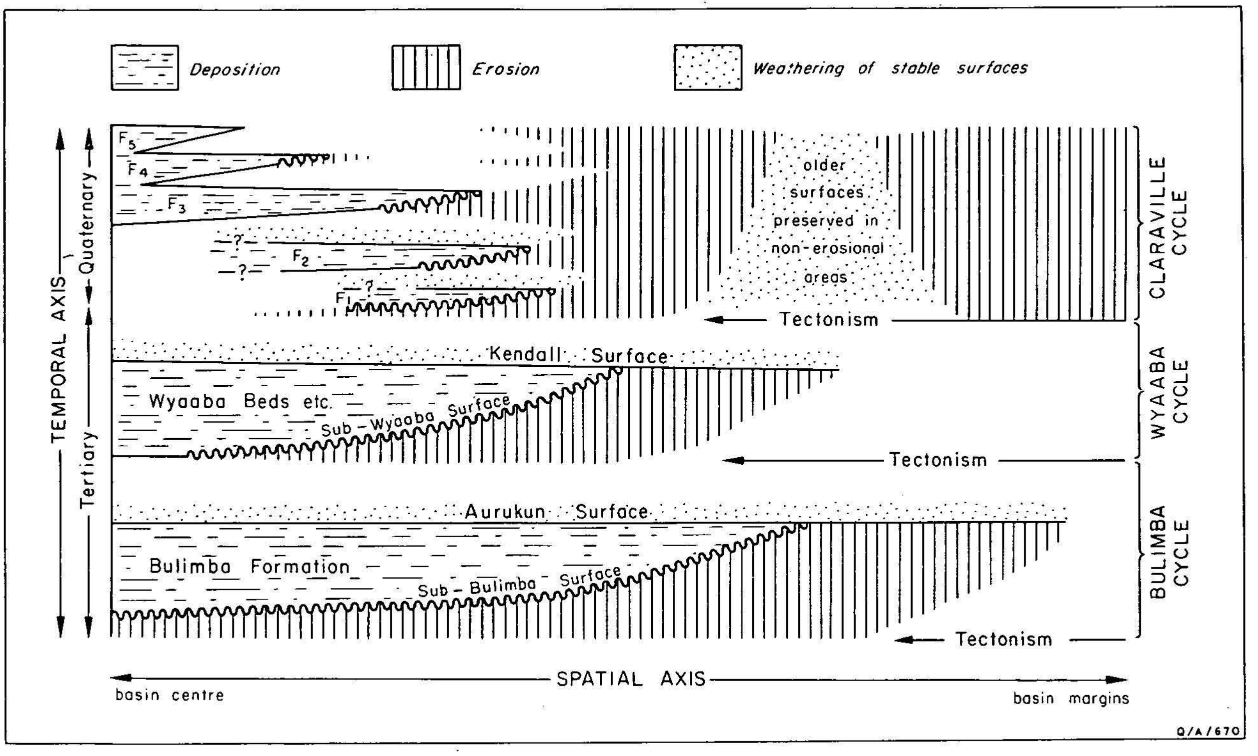


Figure 2. Temporal-spatial relationship diagram showing the relationship between erosional, depositional and weathering events; old land surfaces; and cycles of development in the Karumba Basin. Stages of fan deposition are indicated by F_1, F_2, . . .

Within the region studied, the Sub-Georgina, Sub-Carpentaria, Sub-Bulimba, Sub-Wyaaba and Sub-Claraville Surfaces are all unconformity surfaces. The names are taken from the overlying deposits, or the basins which they floor. Only the Sub-Georgina and Sub-Carpentaria Surfaces have been extensively exhumed. The younger unconformities are generally only recognisable in the subsurface.

The more widespread type of old land surface recognised in the region is the **terminal surface,** which appeared as the relief was reduced towards the end of an active phase of erosion and deposition. The evolution of this type of surface is comparable to that of modern 'peneplains' or 'pediplains', though the genetic distinction between these two types generally cannot be made for ancient surfaces and the general term 'planation surface' is more appropriate. Terminal surfaces can have both erosional and depositional components, the former being most common near the margins of a basin, and the latter in its central parts (cf. Fig. 2). The Aurukun, Tennant Creek, Featherby, Kendall and Campaspe Surfaces are terminal surfaces, though the Campaspe Surface is not as widespread as the others and formed at the end of a subsidiary cycle.

In addition to the types discussed above, several **depositional surfaces** occur on the younger set of fans and alluvial deposits in the area (Grimes & Doutch, 1978). They are only subsidiary breaks within the current Claraville cycle (cf. Fig. 2), and have not been given formal names. The name **Holroyd Surface** was used by Doutch (1976) for a set of desiccation features which he attributed to a period of aridity in late Pleistocene times.

Because the terminal surfaces are of low relief and have survived in the landscape for a considerable time, the rocks beneath them have been subjected to deep weathering and the formation of duricrusts: a factor which has aided in their preservation, and which assists in their recognition and interpretation. Duricrust is used here as a general term for any indurated zone resulting from cementation at or below a land surface as a part of a weathering process. A ferruginous laterite is a duricrust which is cemented by iron; aluminous laterites, some of bauxite grade, are also present. Siliceous duricrusts are impregnated by silica, and when completely silicified are known as silcretes. The cemented duricrust is commonly the upper part of a deep weathering profile. In addition to a thin cover of loose sand at the surface, a laterite profile generally comprises an upper ferruginous or aluminous, cemented zone, the laterite; a middle mottled zone; and, in some places, a lower pallid zone. Ferricrete is used by some authors in the same sense as ferruginous laterite, but here it is used for less well developed ferruginous crusts and iron-cemented weathering detritus which have only a poor profile development, or none at all.

Pre-Cainozoic land surfaces

The pre-Cainozoic geomorphic history is not well known except by deduction from the Mesozoic and older sedimentary rocks (see Smart & others, in prep.). Exhumed remnants of two buried unconformity surfaces are known: the late Precambrian to Cambrian Sub-Georgina Surface, and the Mesozoic Sub-Carpentaria Surface. Both have a fairly high local relief in the areas where they have been exhumed.

The term **'Sub-Georgina Surface'** is suggested here for the 'Pre-Middle Cambrian surface' recognised by Twidale (1966) and de Keyser (1969) in the south-western part of the region (Fig. 1). The non-chronological term is preferred. This is an unconformity surface at the base of the Georgina Basin sequence which has been exhumed in parts of the Isa Uplands. De Keyser & Cook (1972) illustrate a surface which

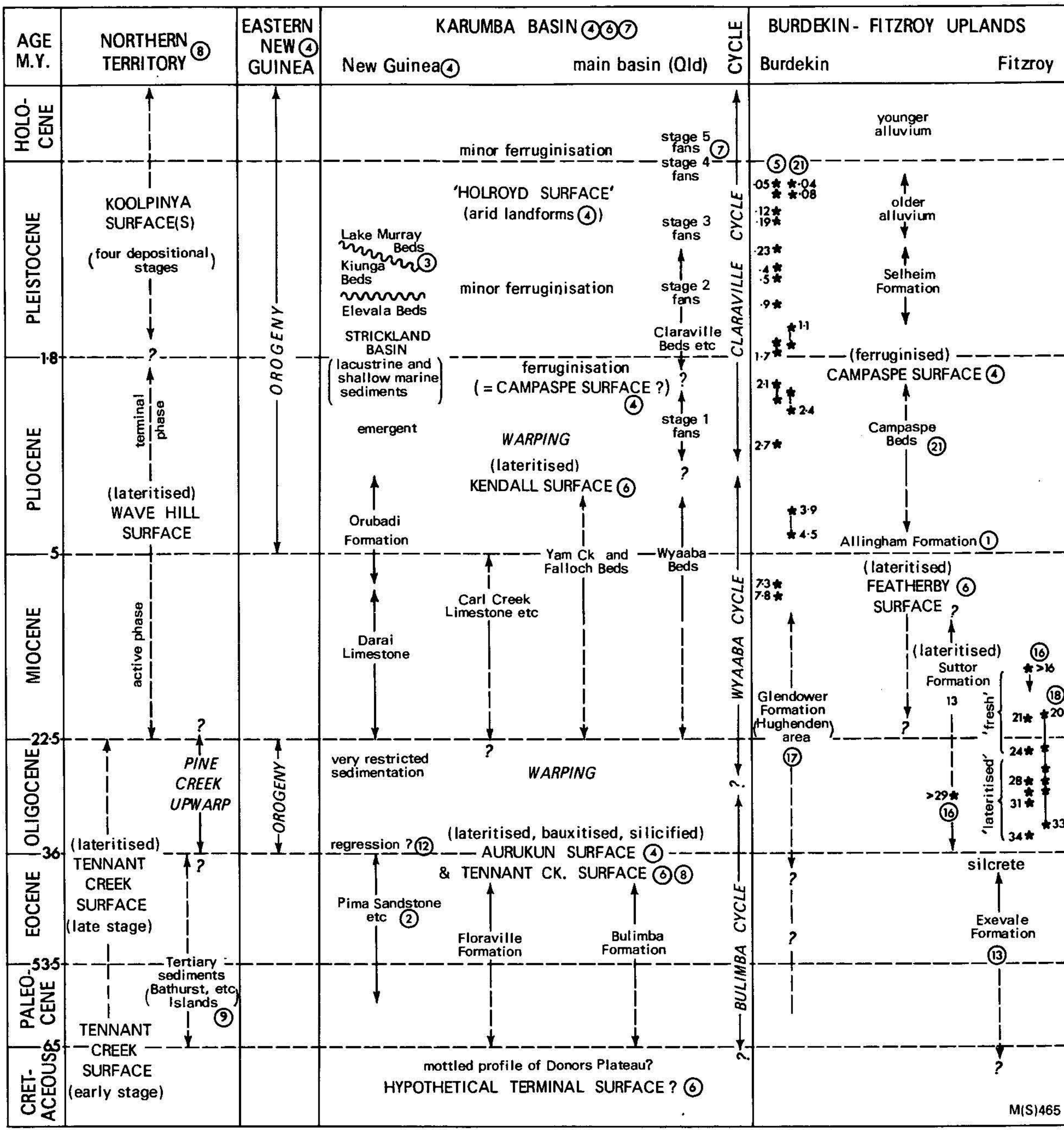
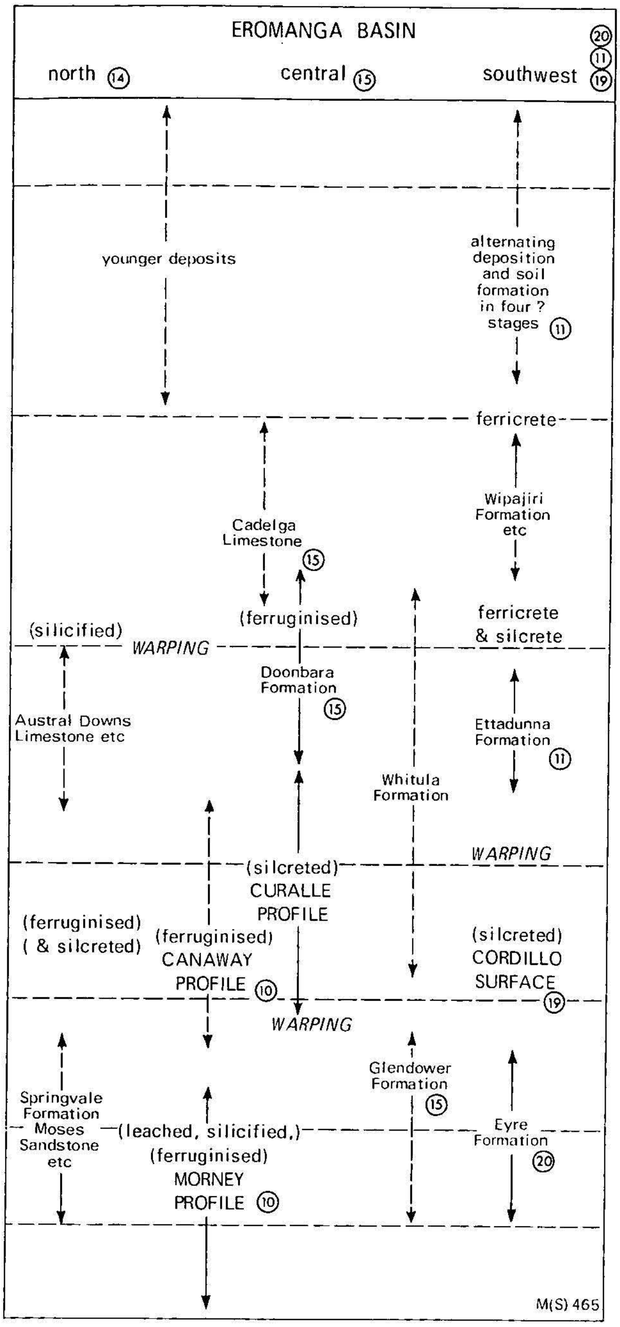


Table 1. Correlation chart for Cainozoic land-surfaces and events in northern Queensland and adjacent regions.
NOTES—*basaltic vulcanism with ages in millions of years. (1) Archer & Wade (1967), (2) Bain & MacKenzie (1974), (3) Blake (1971), (4) Doutch (1976), (5) Griffin & McDougall (1975), (6) Grimes—this paper, (7) Grimes & Doutch (1978), (8) Hays (1967), (9) Hughes & Senior (1974), (10) Idnurm & Senior (1978). (11) Jessup & Norris (1971). (12) Kennett & others (1972). (13) Malone & others (1964), (14) Paten (1964), (15) Senior (1976, 1978), (16) Sutherland, Stubbs & Green (1977), (17) Vine & others (1963), (18) Webb & McDougall (1967), (19) Wopfner (1974), (20) Wopfner & others (1974), (21) Wyatt & Webb (1970).

appears to have had a moderately strong relief. Exhumed remnants of the surface also occur further south (cf. Twidale, 1966, p. 31 & 42).

The **Sub-Carpentaria Surface** is the Jurassic to Early Cretaceous unconformity surface at the base of the Carpentaria Basin sequence. It was buried by the Jurassic and Cretaceous sandstones and the transgressive Wallumbilla Formation and has been exhumed by later erosion in several areas. This is the 'pre-middle Mesozoic Surface' of Twidale (1956, 1966), and the 'Gondwana Surface' of King (1949). The main requirement in the recognition of exhumed parts of this surface is the ability to trace the surface on pre-Mesozoic rocks up to and beneath outcrops of Mesozoic sediments—thereby demonstrating its continuity with the buried unconformity (see Fig. 3 of this paper; and Twidale, 1956, fig. 2). In the absence of Mesozoic deposits the surface can be assigned to an upland surface or summit conformity provided that this lies above the levels of the Tertiary surfaces.

The exhumed surface is best preserved to the southeast of Croydon (Fig. 1) where the Mesozoic sandstones have been stripped from Precambrian volcanics. Here remnants of the sandstone are common in the valleys, well below the level of adjacent hills of the older volcanics. Much of the present drainage pat-



tern and general topography of this area appears to be similar to that which existed in Jurassic times. Twidale (1956, 1966) also recognised the following exhumed and dissected remnants of the surface: the plateau surface of the Newcastle Ranges and other areas of resistant volcanic rocks in the Einasleigh Uplands, the mesas and summit conformities of parts of the Isa Uplands (see also Grimes, 1974), and the benches in the Burke Plain to the south of the Isa Uplands. In the Cloncurry Plain accordant summits dipping towards the Carpentaria Basin appear to be related to a continuation of the buried unconformity (Twidale, 1956; Grimes, 1974). The Sub-Carpentaria may also be present on the dissected high surface of the McIlwrath Plateau in the Peninsula Uplands (Whitaker & Gibson, 1977).

The Sub-Carpentaria Surface has an undulating and in places quite marked local relief; for example, east of Wrotham Park the unconformity beneath the Mesozoic sandstones has abrupt changes in elevation of up to 40 m, resulting from differential erosion of the resistant Palaeozoic volcanics and more easily eroded older sediments. Old gorges and valleys with Mesozoic infillings are also present in the northwestern part of the Isa Highlands (Fig. 3). A seismic reflector which shows a rugged relief beneath the western part of the Gulf is probably a buried part of the Sub-Carpentaria Surface (Smart & others, in prep.).

Deposition in the Carpentaria Basin ended as the sea withdrew before late Cretaceous times. By analogy with the Cainozoic cycles discussed below one would expect a terminal planation surface to have formed as activity in the Carpentaria Basin drew to a close. Such a surface has not been definitely recognised within the study area though de Keyser and Lucas (1968, p. 147) refer to a 'post-Lower Cretaceous' erosional plain which lies to the east of the Carpentaria Basin. The late Cretaceous to early Eocene Morney Profile of Idnurm & Senior (1978) may have developed on a contemporary surface in the Eromanga Basin, and the early stage of the Tennant Creek Surface in the Northern Territory (Hays, 1967) may also be contemporaneous (see Table 1).

Cainozoic land surfaces

The Karumba Basin

The development of the Karumba Basin can be divided into three major cycles, each of which produced deposits and land surfaces (see Introduction and Fig. 2).

The Bulimba Cycle. This first cycle commenced with upwarping of the basin margins in the early Tertiary; it continued into the mid-Tertiary (?Oligocene). In the initial active phase erosion occurred in the uplands and deposition in the depressed areas. These sediments included the continental arkosic sand and clay of the Bulimba Formation on the western side of Cape York Peninsula, and the Floraville Formation to the south of the Gulf of Carpentaria. Both these formations may extend for some distance beneath the present area of the Gulf. As the relief of the area was reduced, a stable planation surface evolved. The surface was lateritised, and in places silicified (the Aurukun Surface).

The **Sub-Bulimba Surface** is the unconformity at the base of the Bulimba Formation and its equivalents (Fig. 4). It has only been locally exhumed. Smart & others (in prep.) show the generalised form of the buried surface beneath the area of the Gilbert-Mitchell Trough. On the basis of its outcrop trace in the Weipa area, and from detailed drilling (Pettifer & others, 1976) it appears to have had a moderately low, undulating, local relief. The Sub-Bulimba Surface was progressively buried by the spreading sediments as the downwarped area was filled. It is therefore a diachronous surface of early Tertiary age, that is it had a similar time span to the Bulimba Formation—which formed with it and buried it (cf. Fig. 2).

The **Aurukun Surface** (Doutch, 1976) is an early to mid-Tertiary planation surface which formed both as a depositional surface on the Bulimba Formation and as an erosional surface on the adjoining areas of older rocks. It is a flat or gently undulating surface which is best preserved on the Weipa Plateau (Smart, 1977) (Fig. 1), where it has a broad pattern of dendritic and commonly swampy valleys. This drainage pattern may have been inherited from the original surface (MacGeehan, 1972), although Doutch (pers. comm.)

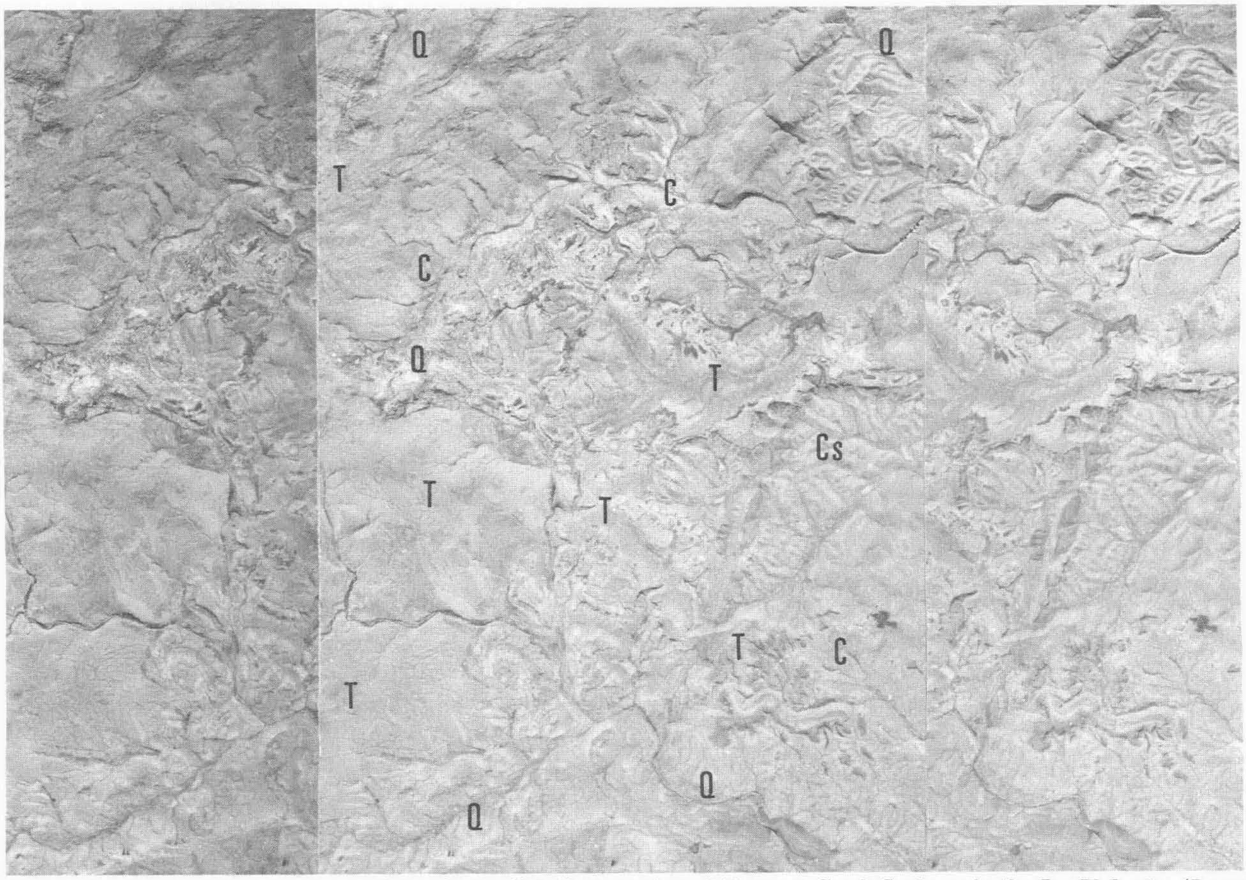


Figure 3. Stereotriplet showing relicts of the Sub-Carpentaria and Tennant Creek Surfaces in the Isa Uplands. (Lawn Hill, CAB 4012, photos 2284, —6, —8). T = Tennant Creek Surface, C = Sub-Carpentaria Surface, Cs = summit levels related to the Sub-Carpentaria Surface, Q = Quaternary erosional areas.

points out that though the valleys pre-date the bauxite they could still be younger than the original Aurukun Surface as the bauxite has developed in several stages.

Deep weathering of the Aurukun Surface has resulted in the formation of well developed laterite and bauxite profiles (Smart & others, in prep.). As later deep-weathering events appear to be superimposed on the mid-Tertiary profile its original nature is difficult to assess. Ferruginous laterites which occur in places buried beneath the Wyaaba Beds in the Gilbert-Mitchell Trough are thought to be part of this profile because of their stratigraphic position, but in the Weipa Plateau, Smart (1977) considers that it was initially an aluminous laterite which has been upgraded to bauxite by later weathering events. Some silicified horizons occur at intervals in the profile in the southern part of the Weipa Plateau and true silcretes have been observed in a few places. Whether this silicification is related to the original or later weathering events is not known. In the Weipa Plateau, the laterite profile is generally capped by about a metre of residual sand, which may represent the A horizon of the profile (Smart, Appendix 4 *in* Doutch & others, 1973). The Kimba Plateau south of Coen, is the only undissected part of the surface

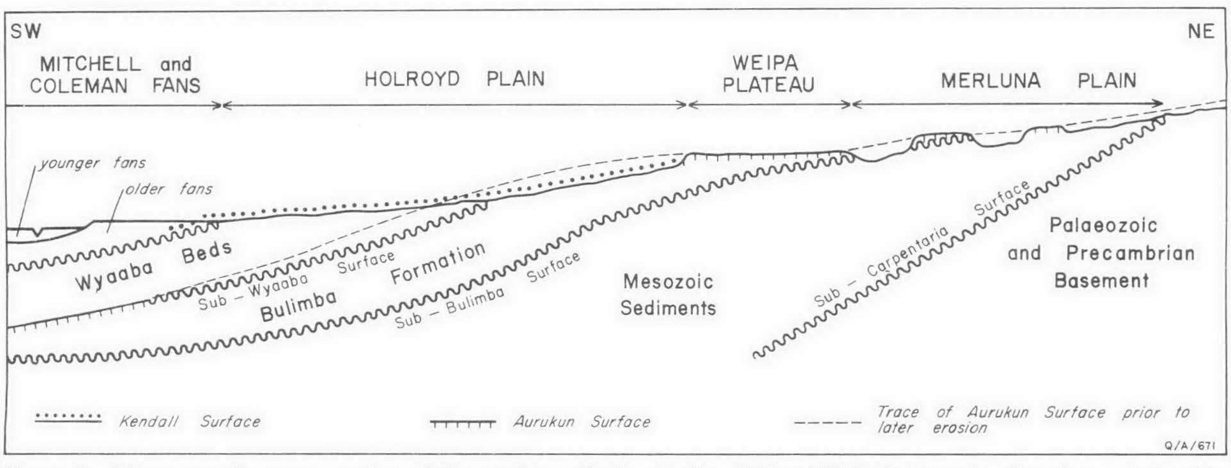


Figure 4. Diagrammatic cross section of the northern flank of the Gilbert-Mitchell Trough. Showing relationships between old land-surfaces and depositional units.

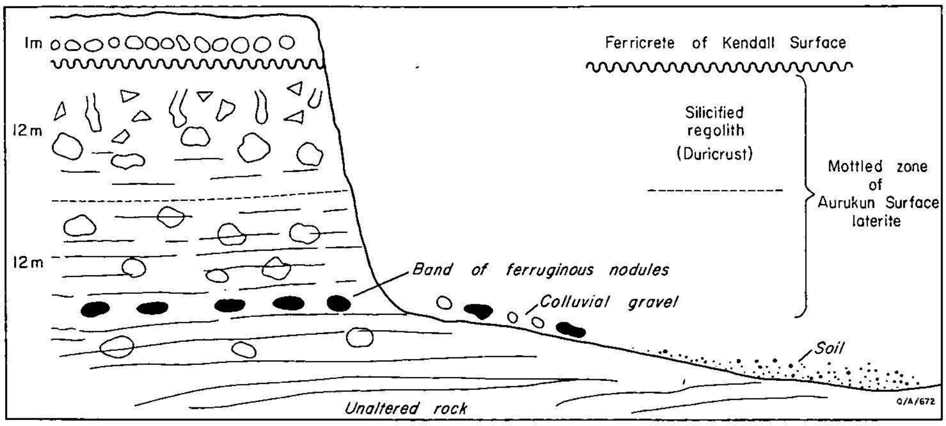


Figure 5. Diagrammatic section of the deep weathering profile on the Donors Plateau (after Doutch & others, 1970).

apart from the Weipa Plateau; ferruginous laterite is developed in this area. Elsewhere in the Karumba Basin erosion has generally removed the upper part of the profile, leaving only the mottled and pallid zones—which have commonly been silicified (Fig. 5). Such areas are the hills and ridges to the east of the Weipa Plateau and, further south, in the Red, Mornington, Normanton, and Donors Plateaus (see Fig. 1). Accordant summits and a few plateau surfaces in the Peninsula Uplands may also be derived from the Aurukun Surface, as some of these have lateritic remnants and they lie above the level of nearby areas of the younger Kendall Surface (cf. fig. 2 of Whitaker & Gibson, 1977).

The age of the Aurukun Surface can at present only be deduced from its position in the overall sequence of development (see Table 1). It is developed on the Bulimba Formation, which is thought to be of early Tertiary age, though this estimate is partly based on ideas concerning the age of the Aurukun Surface (Doutch, 1976; Smart & others, in prep.). The Aurukun Surface pre-dates the Wyaaba Beds, which are probably of Miocene to early Pliocene age. An early to mid-Tertiary age is therefore the best estimate from evidence within the basin.

Doutch (1976) suggested correlations with several early Tertiary surfaces in Australia including the Tennant Creek Surface (Hays, 1967) in the Northern Territory (see below). The termination of the Bulimba Cycle, and therefore of the Aurukun Surface development, could well be related to the Oligocene orogenesis in New Guinea, and warping similar to that in the Karumba Basin occurred elsewhere in Australia in mid-Tertiary times. A mid-Tertiary (Eocene to Oligocene) age is therefore most likely (cf. Table 1).

In summary, the main criteria in the recognition of areas of Aurukun Surface are: firstly its stratigraphic position (illustrated in Fig. 4), and secondly its topographic position above the later Kendall Surface (Fig. 6). The broad drainage pattern and well developed lateritic or bauxitic weathering profile also assist but are not in themselves diagnostic.

The **Tennant Creek Surface** (Hays, 1967) is a Late Cretaceous to early Tertiary planation surface that is well developed in the Northern Territory (Wright, 1963; Hays, 1967). It is correlated here with the surface which caps the Mullaman Beds and the Precambrian rocks of the Isa Uplands in the southwest of the area (see Figs. 1 & 3), i.e. the upland part of the 'Tertiary Surface' of Grimes (1974) and the 'Early to mid-Tertiary Surface' of Twidale (1956, 1966).

In the Isa Uplands the surface is preserved on the top of Mesozoic mesas, and elsewhere as an extensive summit conformity (Fig. 1) (cf. Grimes, 1974, fig. 11). It appears to have been a plain which truncated both the Mesozoic sediments and more resistant Precambrian rocks. Some remnants of old shallow drainage lines are present on the mesa surfaces (Fig. 3). The surface has generally been deeply leached, lateritised and silicified; with thick porcellanites developed on the Mesozoic mudstones.

Hays (1967) deduced a Late Cretaceous to mid-Tertiary age for the Tennant Creek Surface in the Northern Territory, but there seems to have been two stages in its development there (Table 1). The older stage is a laterite developed on Mesozoic rocks and underlying the early Tertiary Van Diemen Sandstone on Bathurst and Melville Islands (Hughes & Senior, 1974; Hays, 1967, p. 198) while the younger stage, which is also lateritised, overlies the Tertiary sediments. In the Isa Highlands the Tennant Creek Surface is developed on the Cretaceous sediments and pre-dates the mid-Miocene Carl Creek Limestone which lies in a valley cut below the level of the surface (Smart & others, in prep.). The surface therefore pre-dates the Wyaaba Cycle, but one cannot be sure which of the two Northern Territory Stages (if not both) is represented in the Isa Uplands. On Table 1 the surface is tentatively correlated with the younger of the two stages, and with the Aurukun Surface.

The term **Strathgordon Surface** was first used, without definition, by Powell & others (1976) for a (ferruginous) surface developed on the Wyaaba, Lilyvale, Yam Creek and Falloch Beds, which they considered to be contemporaneous units. Doutch (1976) considered, on the basis of denudation chronologies in the separate areas, that the Wyaaba Beds were older than the other units. He used the name Strathgordon Surface for a set of silicified surfaces within the Karumba Basin which he thought postdated the Wyaaba Beds. He correlated the ferruginous surface on the other units with the Campaspe Surface (see below). Doutch's description of the Strathgordon Surface is brief and no reference area is given. It contains a conflict between the criteria of silicification and of post-Wyaaba age, and further by his apparent application of the term to ferruginised surfaces on the Wyaaba Beds and elsewhere. Doutch (pers. comm.) has indicated that he considers the criteria of silicification more important than the concept of a post-Wyaaba age in defining the Strathgordon Surface.

The main siliceous surfaces in the Karumba Basin are older than the Wyaaba Beds and are either equivalent to the Aurukun Surface or represent a late stage in its development. To avoid the confusion resulting from the different applications of the term 'Strathgordon', the name Kendall Surface will be applied below to the ferruginised terminal surface on the Wyaaba Beds.

'Strathgordon' silicification in the Karumba Basin is best developed on the Tennant Creek Surface on the Isa Uplands, and in conjunction with mottled and pallid zones of the Aurukun Surface in the Donors and Red Plateaus, though most of the surfaces in the region have localised areas of minor silicification.

Doutch & others (1970) described a profile in the Donors Plateau which consists of a ferricrete zone (commonly stripped) and a mottled zone which has been silicified in its upper part to form a duricrust (Fig. 5). They attributed this profile to three stages of deep weathering: an initial mottling as part of a laterite profile, a second stage in which the mottled zone was silicified after stripping of the upper part of the profile, and a final stage in which the ferricrete zone was formed. Doutch (pers. comm.) would correlate these three stages with the Aurukun Surface, the 'Strathgordon Surface' (in his sense), and the Kendall Surface (as used here) respectively.

On the Donors Plateau the 'Strathgordon' duricrust produces a pattern of old valley forms which are now in inverted relief as a result of the erosion of their presumably less indurated interfluves. Similar valley forms in inverted relief occur in the Red Plateau, north of Georgetown. There, some of the duricrusted valleys contain a younger fill which is correlated with the Wyaaba Beds as in places it can be seen continuing beneath the Kendall Surface at the margins of the Holroyd Plain, and Wyaaba Beds underlie this surface.

Doutch's 'Strathgordon Surface' would therefore appear to have developed prior to the deposition of the Wyaaba Beds. If the underlying mottled profile in the Donors Plateau is assigned to the Aurukun Surface, then the relationship suggests that the 'Strathgordon Surface' is a late stage of the Aurukun Surface in which siliceous solutions leached from the interfluves were deposited in the adjoining valleys. On the other hand if the mottled material were considered to be older, and perhaps could be correlated with the Morney profile of Senior (1976), then the Strathgordon Surface would be a simple correlate of the Aurukun Surface.

Silcretes and siliceous duricrusts become more common and better developed towards the south. No silcretes have been identified north of about 13°S latitude. On the other hand the Tennant Creek Surface on the Isa Uplands is extensively silicified. This trend could represent a climatically controlled transition within the weathering profiles of the time; from dominantly lateritic in the north to dominantly siliceous in the inland parts. Lithology exerts some local control but the regional trend is seen on all the lithologies.

In view of the differing uses to which the term 'Strathgordon Surface' has been put, and the suggestion here that if applied as a siliceous surface it is mainly a chemically distinctive 'facies' of the Aurukun Surface, it is recommended that the term 'Strathgordon Surface' be discontinued.

The Wyaaba Cycle. This cycle commenced in the late Oligocene as a result of renewed upwarping of the Karumba Basin margins and downwarping of the Gil-

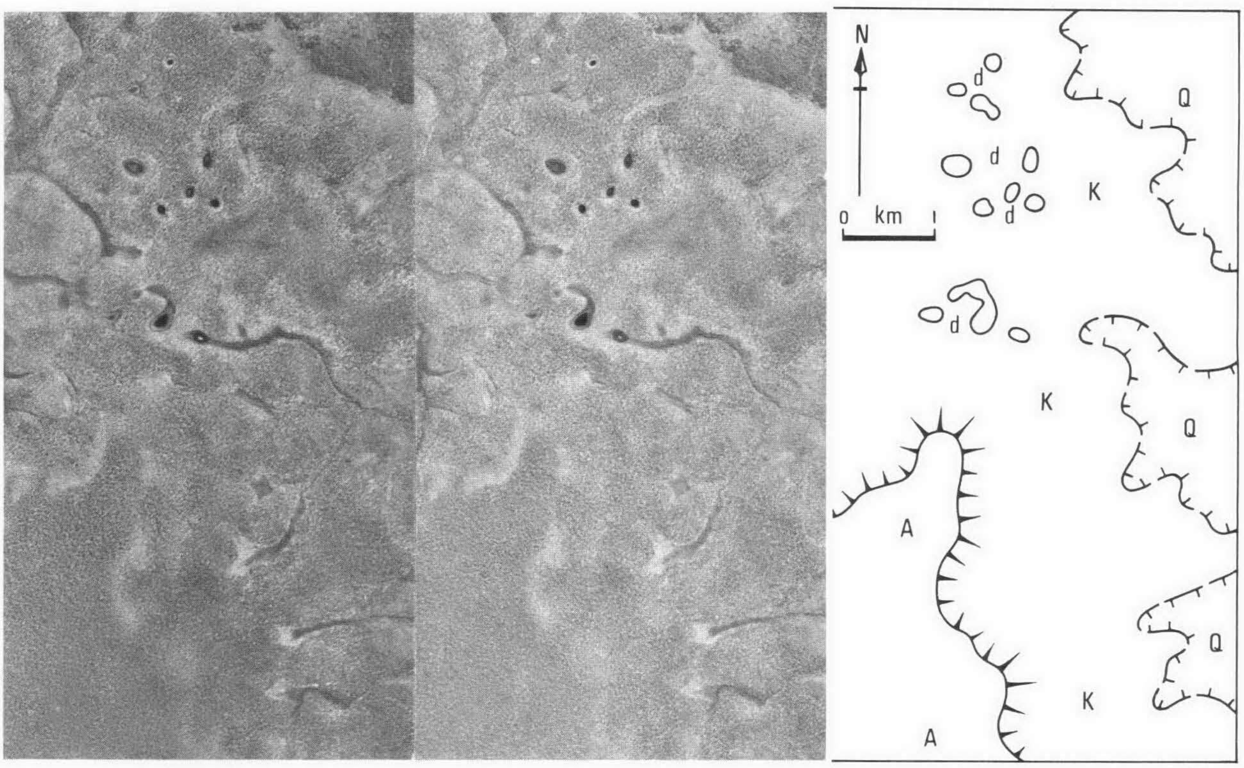


Figure 6. Stereopair showing Aurukun (A) and Kendall (K) Surfaces, and more recent erosion (Q) at the south-eastern margin of the Weipa Plateau (Ebagoola CAB 358, Run 2, photos 66, 68). d = closed depressions on the Kendall surface which Doutch (1976) considers to be deflation features of the Holroyd Surface, but which this author attributes to a laterite-karst effect.

bert-Mitchell Trough. The initiation of the cycle may be related to an Oligocene orogeny at the northern edge of the Australian Plate in New Guinea. The active phase disrupted the development of the Aurukun and Tennant Creek Surfaces. Dominantly continental deposition of a similar type to that of the preceding cycle occurred in the Gilbert-Mitchell Trough (Wyaaba Beds), and in smaller isolated basins (the Yam Creek and Falloch Beds in Cape York Peninsula, and the Lilyvale Beds of the Laura Basin). The sea appeared in the Gulf area for a brief period during deposition of the Wyaaba Beds. Springs deposited the Carl Creek and Gregory Downs Limestones in the southwest of the Karumba Basin. As the cycle's active phase drew to a close the Kendall Surface developed and was lateritised or ferricreted. The Featherby Surface of the Charters Towers area also formed at the end of this cycle, though in that area stability would seem to have been achieved earlier than in the Karumba Basin. Both surfaces were disrupted and the cycle terminated by uplift, with associated vulcanism, in the Pliocene.

The **Sub-Wyaaba Surface** is the unconformity surface which has been recognised beneath the Wyaaba Beds in the Gilbert-Mitchell Trough (Fig. 4). No exhumed equivalents have been recognised, as the outcrop trace of the surface is mantled by younger colluvial sands. Mottled material has been identified in drill holes from immediately beneath the unconformity surface. This is probably a partly truncated remnant of the earlier Aurukun Surface weathering profile (Fig. 4). Equivalent unconformity surfaces exist beneath the Lilyvale, Falloch, and Yam Creek Beds (Whitaker & Gibson, 1977).

The name **Kendall Surface** is applied here to the planation surface which is preserved in the interfluves of the Holroyd, Croydon, and Strathpark Plains (Fig. 1). This is the surface which was originally referred to (without definition) as the 'Strathgordon Surface' (Powell & others, 1976), but, as discussed as above, that name is best discarded. In the Explanatory Notes accompanying the geological maps of Cape York Peninsula, the Kendall Surface, as used here, was generally referred to as the 'Campaspe Surface', following Doutch's (1976) correlations. The latter term is best restricted to the Burdekin Uplands as its correlation with the Karumba Basin surfaces is uncertain (see below).

The name Kendall Surface is derived from the Kendall River and the northern part of the Holroyd Plain between the Kendall and Holroyd Rivers is suggested as a reference area. The surface is widespread throughout the Karumba Basin (see Fig. 1). On the Holroyd Plain it has a characteristic dendritic drainage pattern of shallow swampy valleys, which is illustrated in Doutch & others (1971, fig. 3).

The Kendall Surface developed towards the end of the Wyaaba Cycle as a depositional surface on the Wyaaba Beds around the Gilbert-Mitchell Trough; on the Lilyvale, Falloch, and Yam Creek Beds in the Laura Basin and Cape York Peninsula; and as an erosional surface developed at the expense of the Aurukun Surface on parts of the Bulimba Formation and older rocks elsewhere (Fig. 4). The depositional surface has undergone some mild erosional modification since its formation. The surface has been lateritised and ferricreted, and is also silicified in some places. The laterites are best exposed along the western side of the Lynd River and in the Strathpark Plain. The Yam Creek Beds have ferricrete cappings up to 3 m thick.

Part of the Doomadgee Plain may be a correlative of the Kendall Surface (Fig. 1). Here the laterites are very well developed. In the Laura Basin the Jack Plain may also be equivalent; it has been poorly lateritised and, in places, silicified.

The Kendall Surface is of Pliocene age as it post-dates the Wyaaba Beds, of Miocene to Pliocene age (Smart & others, in prep.), and pre-dates the late Pliocene to Quaternary fan deposits (Grimes & Doutch, 1978). The Gilberton Plateau was upwarped during the Pliocene (Smart & others, in prep.); the duricrusted surface on the plateau could be an upwarped equivalent of the Kendall Surface on the Strathpark Plain to the west (Smart, 1973). This plateau surface is an extension of the Featherby Surface of the Burdekin Uplands. If the correlation with the Strathpark Plain is valid then the overall Kendall-Featherby surface must be diachronous, as to the southeast the Featherby Surface is older than Pliocene. In view of the different ages, and the lack of positive proof of continuity, a separate terminology is adopted for the two areas.

The main criteria for the recognition of the Kendall Surface are firstly its stratigraphic position: younger than the Wyaaba Beds and Aurukun Surface, but older than the current cycle of erosion and deposition; secondly its topographic position below the level of the Aurukun Surface; and to a lesser extent its typical landform in the Holroyd Plain, and the nature of its laterite profile which is generally not as strongly developed as the Aurukun Surface.

The Claraville Cycle. Pliocene tectonism upwarped the Gilberton Plateau and other areas, and initiated the Claraville cycle. This cycle has continued to the present; it contains a number of recognisable subcycles, for instance the five stages of fan development in the Gilbert-Mitchell area (Grimes & Doutch, 1978). Contemporaneous volcanism occurred in the Burdekin Uplands.

The **Sub-Claraville Surface** is an unconformity surface beneath the deposits of the present Claraville Cycle. It has not been exhumed. The lithological similarities between the fans and the older Wyaaba Beds make the surface difficult to identify in some areas. It has only been mapped south of the Gilbert-Mitchell Trough, where the deposits overlie Cretaceous rocks (Grimes & Doutch, 1978, fig. 5).

Grimes & Doutch (1978) have recognised five stages of deposition of fans and related alluvial plains in the Karumba Basin. Each of these stages has left a depositional surface, and the recognition of the stages is based to a large extent on the morphology and relationships of these surfaces.

Within the Gilbert and Mitchell fans area the oldest, stage 1, fans are generally restricted to valley sides in the fan heads. Further to the south and west, stage 1 plains stand topographically above the younger surfaces; they have been ferruginised. Stage 2 fans are also ferruginised. Stage 3 fans are widespread and may have formed during the last interglacial. Stage 4 and 5 are late Pleistocene and Holocene flood plains.

Doutch (1976) applied the name **Holroyd Surface** to a suite of features developed on Stage 3 fans and most older surfaces, which he considers formed during a period of desiccation. The concept is therefore different to that of the other surfaces discussed here. Doutch cited 'sand dunes on the Millungera Plain and dune-like forms, sand plains with clay pans, deflation features and choked drainage' as evidence for aridity. He considered that these features occurred after the formation of the

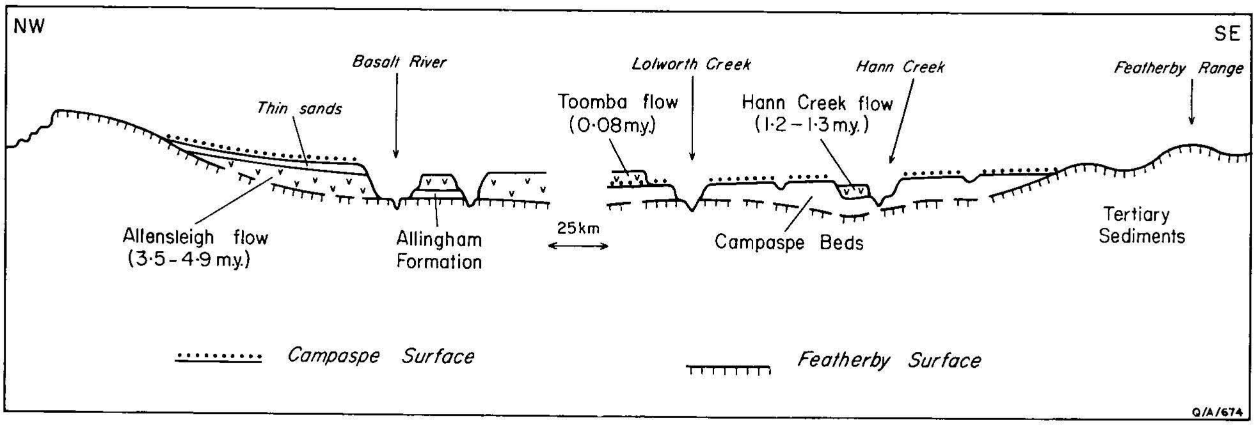


Figure 7. Diagrammatic cross section north-west of Charters Towers, showing the relationship between the old land surfaces and dated basalt flows.

Stage 3 fans and the oldest beach ridges (100 000-120 000 years B.P.) and before the formation of the stage 4 fans and the younger, Holocene ridges (6500 years B.P. or older). They may correlate with the end of the last glacial maximum, which was a time of lower rainfall (Kershaw, 1975).

Although I do not dispute that there was a period (or periods) of aridity during the last glaciation, I doubt if all of the features cited by Doutch should be attributed to arid effects. Some of the 'clay pans', which Doutch apparently considers to be deflation hollows, could also be laterite-karst depressions of the type described on the Doomadgee Plain (Grimes, 1974), or even true karst in the case of the leached beach ridges. Most of the sand plains in the Karumba Basin lie above lateritised surfaces; these sands probably represent the A horizons of the laterite profiles (cf. Smart, appendix 4 *in* Doutch & others, 1973); that is, they indicate a seasonal humid climate, rather than aridity. By 'dune-like forms' Doutch (1976) appears to be referring to the wanderrie patterns described by Grimes & Doutch (1978). These occur only on the Stage 1 and 2 fans, and are truncated by the earliest of the Stage 3 fans. On the other hand the 'clay pans' and the dunes of the Millungera Plain are younger, because they occur on the stage 3 fans and the oldest beach ridges. This suggests at least two periods of aridity, whereas Doutch's usage of the term Holroyd Surface implies a single event.

Land-surfaces of the Burdekin Uplands

The early Tertiary history of the Burdekin Uplands is not well known. The Glendower Formation, though considered to be of Pliocene age by Whitehouse (1940), may well be older and related to the Wyaaba Cycle—or possibly even the Bulimba Cycle (cf. Table 1). The nature of its contained silcrete boulders has been debated by Vine & others (1963) who consider that they may have formed *in situ*, and not been derived from an older silcreted surface as was suggested by Whitehouse. However, the conglomerates in the formation contain silcrete cobbles which appear to be definitely detrital. The Glendower Formation would seem to have formed after a period of silicification; the Formation has subsequently been lateritised, and locally silicified—possibly in separate events. Tertiary sediments, the Southern Cross Formation of Wyatt (1970), occur beneath a lateritised surface in the Charters Towers area, and may be of similar age.

The name **Featherby Surface** is applied here to an undulating lateritised surface in the Burdekin and Einasleigh Uplands (Figs. 7 & 8). The surface was first mapped by Wyatt & others (1970, 1971) as laterite (unit Tl) on the Charters Towers and Townsville 1:250 000 Geological Sheets. A suitable reference area for the surface is on, and west of, the Featherby Range, about 18 km west of Charters Towers. It would appear to have been a stable surface which was lateritised at the end of the Wyaaba Cycle. Wyatt & others (1970, p. 63) described the well-developed laterite profile, and they also related the silcretes in the area to this surface. The Featherby Surface is well developed on the Tertiary sediments and older rocks in the Charters Towers-Hughenden area and further south. Farther to the north and west it is generally only represented by isolated mesas (Fig. 1), which hinders its extrapolation into the Karumba Basin. However, these remnants can be traced as far as the Gilberton Plateau and the Lucy Tableland (Fig. 8). The older parts of the Alice Tableland to the south (Whitehouse, 1940; Vine & Doutch, 1972) would also appear to be correlates of the Featherby Surface; as would the gently undulating surface of the 'Denna Plain' (Coventry, in prep.) and the surface on the Glendower Formation, both areas lying east of Hughenden.

Although typified by thick laterite profiles and red-earth soils, parts of the surface also exhibit yellow earths (Coventry, in prep.). The undulating nature of the surface on the Southern Cross Formation indicates that there has been some erosion since the deposition of that unit. It may be that there has been more than one stage of weathering in the evolution of this surface, with the yellow earths formed by partial stripping and reweathering of the initial laterite profile. Some thick, uniform, red earths found in the lower parts of the surface may be redistributed lateritic soils derived from the interfluves.

The Featherby Surface can be traced beneath the Campaspe Beds (Fig. 7), and buried laterites of the surface are exposed at Red Falls (Wyatt & others, 1970; p. 63 and figs. 16 & 17). The surface also occurs beneath basalt flows of the Nulla Province; slightly stripped laterite profiles of the surface underlie both the Allensleigh Flow (dated at 4.5 m.y. by Wyatt & Webb, 1970) and the early Pliocene Allingham Formation: near Allensleigh and Bluff Downs homesteads respectively (Wyatt, 1969; Archer & Wade, 1976). Thus the Featherby Surface has an upper age limit of earliest Pliocene.

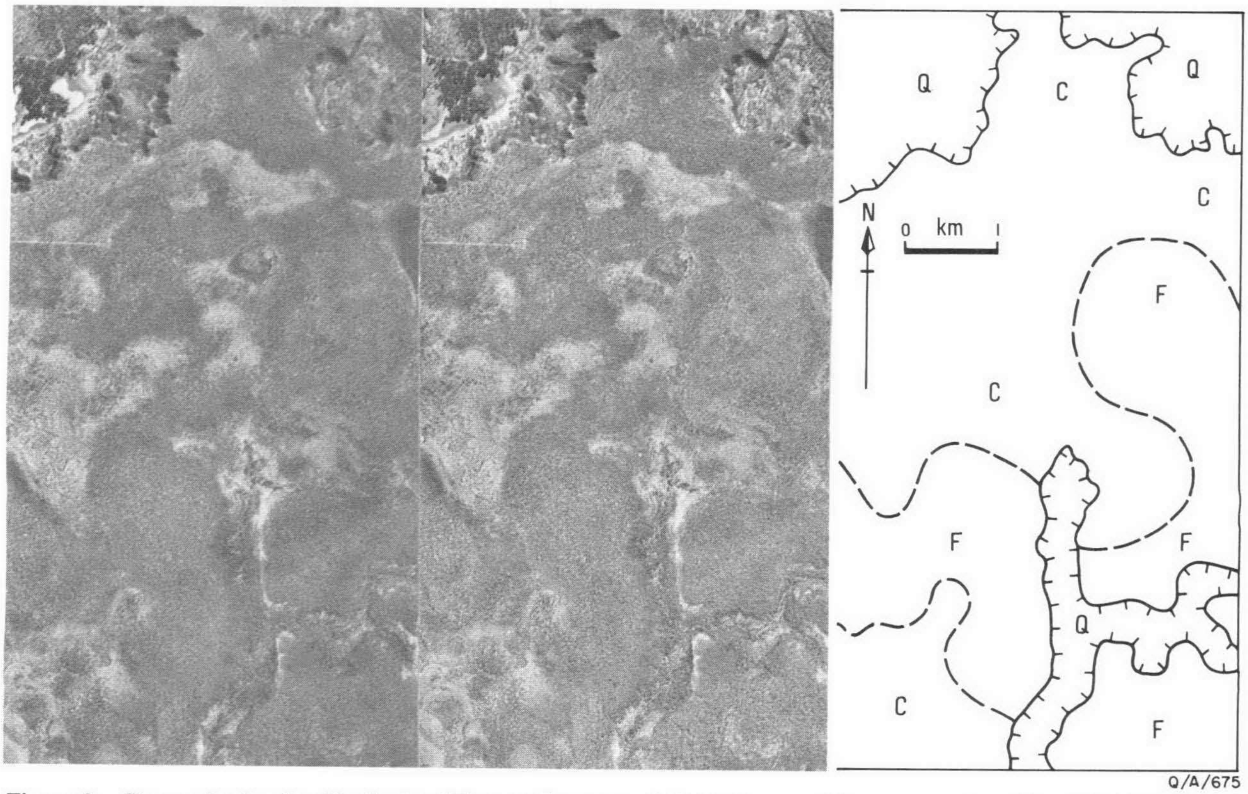


Figure 8. Stereopair showing Featherby (F) and Campaspe (C) Surfaces, and more recent erosion (Q) in the Lucy Tableland, south of Mount Garnet. (Einasleigh CAB 278, Run 5, photos 30, 32).

Remnants of the surface can be traced on air photos into the Mount Coolon 1:250 000 Sheet area, which is southeast of the area shown on Fig. 1. On the Mount Coolon Sheet a possible lower age limit can be deduced. Here the surface is developed on the Suttor Formation, which in turn overlies a lateritised basalt flow with a minimum age of least 29.2 m.y. (Sutherland & others, 1977). If the correlation between the two areas is valid then the age of the Featherby Surface could range from Oligocene to earliest Pliocene, with a Miocene age most likely.

The Featherby Surface can be recognised by its stratigraphic and topographic position (Figs. 7 & 8): it either stands above the younger Campaspe Surface, or is buried by the Campaspe Beds and their equivalents. The undulating surface is distinct from the flat to gently sloping Campaspe Surface. The laterite is generally better developed on the Featherby Surface.

The early Pliocene upwarping heralded the start of the current Claraville Cycle. The earliest deposits in the Burdekin Uplands are the Campaspe Beds (Wyatt & others, 1970). The **Campaspe Surface** is the flat-lying depositional surface on these beds in the Charters Towers area. It has a ferricreted surface which Wyatt & others (1970, 1971) mapped as unit Tf. Doutch (1976) first applied the name Campaspe Surface, and correlated it with the surfaces on the Yam Creek and Falloch Beds in the Karumba Basin. The term was also applied to what is here called the Kendall Surface (see above).

The Campaspe Surface is best developed on the Campaspe Beds northwest of Charters Towers. There it is a depositional surface, but some erosional equivalents can be recognised elsewhere in the region (see below). Wyatt & others (1970, p. 63) describe the poorly developed laterite profile, which consists of less than a metre of ferricrete overlying a thin mottled zone. This profile is not as well developed as the laterite of the older Featherby Surface, although in the Lucy Tableland (Fig. 8) an equivalent of the Campaspe Surface has a better developed profile, which is difficult to distinguish from adjoining areas of Featherby Surface. The Campaspe Surface is well preserved in the Charters Towers area, generally with only minor dissection near streams. In places it is buried by deposits that are apparently equivalent to the Stage 3 and younger fans of the Karumba Basin (cf. Grimes & Doutch, 1978). In the Einasleigh Uplands there are erosional and depositional surfaces which lie below the level of the remnants of the Featherby Surface, but pre-date later erosion; they are therefore correlated with the Campaspe Surface.

The age of the Campaspe Surface can be deduced in the Charters Towers area (Fig. 7). Wyatt (1969) reported that ferricretes, probably part of the Campaspe Surface, are developed on a thin outwash deposit covering the Allensleigh Flow of the Nulla Basalt. Wyatt & Webb (1970, p. 47) suggest that the Campaspe Beds, which underlie the Campaspe Surface, may also postdate the Allensleigh Flow, which has an age between 3.9 and 4.5 m.y., though they admit to some uncertainty. Near Wyandotte homestead in the Einasleigh 1:250 000 Sheet area, a 2.3 m.y. basalt flow (Griffin & McDougall, 1976) antedates an erosional surface which might correlate with the Campaspe Surface. The Campaspe Surface antedates the Hann Creek Flow (1.2-1.3 m.y.) which lies in a valley cut into the surface (Fig. 7) (Wyatt & others, 1970, p. 66). The age of the surface may therefore be between 3.9 and 1.3 m.y.—possibly between 2.3 and 1.3 m.y.; that is mid-Pliocene to early Pleistocene.

Conclusions

This paper defines and describes a number of old land surfaces in northern Queensland. Many of these had already been recognised by earlier workers (e.g. Twidale, 1956, 1966; Wyatt & Webb, 1970) but had not been formally named. Some age names had been used, but I consider this inadvisable because of the lack of definite data on the ages of the surfaces.

Usage of the term 'Strathgordon Surface' has become confused, and this term should be abandoned.

The recent geological mapping in the Carpentaria and Karumba Basins, together with correlations with adjoining regions, has allowed the surfaces to be placed in the overall stratigraphic sequence. The regional correlations of the old land surfaces were complicated by differences in the local tectonic histories in different areas, and by variations in the nature of the weathering profiles. The tectonic variations caused differences in the times at which the periods of stability and activity started and ended. The variations in profile are probably the result of regional differences in the climates of the time.

The old land surfaces were recognised in the first place by their topographic form and their soils—which contrast with the currently forming surface. Particular land surfaces were then identified and correlated on the basis of their stratigraphic and topographic relationships, and to a lesser extent by their surface morphology and the nature of the duricrusts or weathering profiles developed on them. In the case of the exhumed unconformity surfaces the main criteria was the ability to trace the surface back to the unconformity and thereby establish its stratigraphic position (cf. Fig. 3). Where this was not possible then the topographic position in places allowed the surface to be identified.

The terminal surfaces were identified by their stratigraphic relationships to each other and to the related deposits. These relationships are illustrated in Figs. 2, 4, & 7. In addition, their relative topographic positions helped in the recognition of the sequences. The younger surfaces either lie at a lower level and evolved at the expense of the older, higher surfaces (Fig. 6), or formed on deposits which have buried the older surfaces (Fig. 8). Most of the terminal surfaces have a similar form, but the Kendall Surface sometimes has a distinctive drainage pattern, while the undulating nature of the Featherby Surface is useful in distinguishing it from the flatter Campaspe Surface in the Burdekin Uplands. The duricrusts and weathering profiles, while diagnostic of the terminal surfaces as a group, were less useful in distinguishing or correlating particular surfaces, as the degree of weathering and its nature shows regional variations resulting from climatic trends, and local variations caused by lithological differences. The laterites could provide material suitable for palaeomagnetic dating.

The oldest surfaces within the region are the Sub-Georgina and Sub-Carpentaria Surfaces. These are both preserved as unconformities beneath a sediment cover, and have been partly exhumed. The oldest terminal surface expected in the area would be that formed at the end of the Carpentaria Basin cycle, but within the region this appears to have been completely destroyed by later erosion. However, in the Northern Territory it could be represented by the early stage of the Tennant Creek Surface (Hays, 1967); and to the south of the region the late Cretaceous to early Eocene Morney Profile of Idnurm & Senior (1978) is developed on an analogous terminal surface in the Eromanga Basin (see Table 1).

The geological evolution of the region in the Cainozoic is summarised in Table 1. The Bulimba Cycle was initiated by earth movements in late Cretaceous or early Tertiary time. The Sub-Bulimba Surface is the unconformity buried beneath the early Tertiary sediments of this cycle. The terminal Aurukun Surface appeared towards the end of the cycle, and has been deeply weathered to form aluminous and ferruginous laterites, with some silicification in the south. Possible equivalents outside the Karumba Basin are the later stage of the Tennant Creek Surface (Hays, 1967) in the Northern Territory, which extends as a ferruginised and silicified surface into the Isa Uplands, and the silicified and ferruginised Curalle and Canaway Profiles developed on the Cordillo Surface of the Eromanga Basin (Wopfner, 1974; Idnurm & Senior, 1978). The Canaway Profile has been dated as late Eocene to early Miocene (Idnurm & Senior, 1978). The pre-basaltic silcretes of the Fitzroy uplands may also be contemporaneous (Table 1) with this profile.

The Wyaaba Cycle started with Oligocene warping, and continued until early Pliocene times. The terminal Kendall Surface of this cycle has a laterite profile developed on it. It is of Pliocene age and may correlate with the Wave Hill Surface of Hays (1967) in the Northern Territory. The Featherby Surface in the Burdekin Uplands occupies a similar stratigraphic position, but appears to have been terminated earlier by renewed earth movements in the late Miocene and Pliocene. Correlations with the Eromanga Basin are uncertain at present, though some late Miocene to Pliocene silcretes and ferricretes do occur in that area (Table 1).

The Campaspe Cycle started in the Pliocene, and has continued to the present. The Kendall Surface may have been continuous with the Featherby Surface of the upwarped Gilberton Plateau, and if one compares the local denudational sequences then the relationship between the Featherby Surface, Campaspe Beds, Campaspe Surface, and younger alluvial deposits in the Charters Towers area is similar to that between the Kendall Surface and the fans units in the Carpentaria Plains (Table 1). If compared in this way the Kendall Surface occupies a similar position in the sequence to the Featherby Surface, and the Campaspe Beds and Campaspe Surface might be equivalent to the oldest fans, Stage 1 in the nomenclature of Grimes & Doutch (1978). As discussed above the oldest fans surfaces in the Karumba Basin have been partly ferruginised and silicified. In the Northern Territory the Koolpinya surfaces formed in several stages during this cycle (Hays, 1967). The Holroyd Surface of Doutch (1976) differs in concept from the other surfaces in the region. The term was used for a set of climatically controlled landforms.

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Magnetostratigraphic tests of lithostratigraphic correlations between latest Proterozoic sequences in the Ngalia, Georgina and Amadeus Basins, central Australia

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Palaeomagnetic polarity sequences have been determined for the lower Yuendumu Sandstone of the Ngalia Basin, and the upper Wonnadinna Dolomite and lower Gnallan-a-gea Arkose of the Georgina Basin to illustrate an application of magnetostratigraphy, i.e. to test previous correlations.

The Yuendumu Sandstone of the Ngalia Basin had been correlated with the Arumbera Sandstone because of lithological similarities, as well as palaeogeographic and tectonic considerations. The discovery of trace fossils in the upper Yuendumu Sandstone had supported a correlation with the upper, Cambrian, part of the Arumbera Sandstone. The age of the lower Yuendumu Sandstone was unknown. It was found to be of mixed polarity with a marked normal bias. Magnetostratigraphic comparisons with the polarity pattern from the Amadeus Basin are equivocal: they allow correlations with both the upper part of Arumbera Sandstone I (latest Proterozoic), and the Arumbera Sandstone II-III (Early Cambrian). Since Arumbera II lies unconformably on Arumbera I, we searched for a break within the Yuendumu Sandstone. Photointerpretation and field observation revealed an unconformity within the formation, which, together with the palaeomagnetic data, indicates that the lower part of the sandstone is best correlated with the upper part of Arumbera Sandstone I, and thus is of latest Proterozoic age.

The polarity pattern from the Georgina Basin stimulated a reassessment of possible correlations and the collection of new field data. The magnetostratigraphic test was to check whether the reversely magnetised Julie Formation of the Amadeus Basin correlated with the Wonnadinna Dolomite of the Georgina Basin. The magnetostratigraphic result was negative; the Wonnadinna Dolomite is predominantly normally magnetised. This led to an examination of drill cores from near Mount Skinner on the Alcoota 1:250 000 Sheet area, which showed that the Grant Bluff Formation, which elsewhere overlies the Gnallan-a-gea Arkose and the Wonnadinna Dolomite, occurs at depth below correlatives of Arumbera I, the Julie Formation and the upper Pertatataka Formation. It now appears that the Grant Bluff Formation and the Gnallan-a-gea Arkose (Georgina Basin) are correlatives of the Cyclops and Waldo Pedlar Members of the Pertatataka Formation (Amadeus Basin). The Wonnadinna Dolomite is thought to be a correlative of the 'cap dolomites' above the upper tillites of the Ngalia and Amadeus Basins.

The new Adelaidean stratigraphy of the central Australian basins has immediate application with regard to Adelaidean Polarity Time Scales; it allows for partial coverage back to the time of the upper glaciation of the late Proterozoic.

Introduction

Magnetostratigraphy offers considerable promise as a powerful method of time-correlation, especially with Proterozoic rocks where most other available methods are much less precise. Stratigraphers still regard the method with much scepticism, as is proper considering the many problems in rock magnetism, and the general paucity of data. But for clearly defined correlation problems, the method has immediate application. Recent stratigraphic studies by the BMR in the Ngalia and Georgina Basins have produced new and more precise correlations between the three Adelaidean (late Proterozoic) basins of central Australia (Fig. 1) and in addition, the continuing close international interest in the Proterozoic-Cambrian transition has created a demand for more information relating to that interval. It was against that background that the first author accompanied a combined BMR-Geological Survey of South Australia group which critically examined the lithological correlations of Proterozoic tillites in central Australia (Preiss & others, 1978). During that trip, several stratigraphic sections were selected for magnetostratigraphic studies; some results of this work are described here.

The magnetostratigraphic correlations described here are made possible by the work of Kirschvink (1978) in the Amadeus Basin. His work provides a magnetostratigraphic reference for the latest Adelaidean and earliest Cambrian. The work reported here in both the Ngalia and Georgina Basins led to a re-examination of the lithostratigraphic correlations in this interval. In the Ngalia Basin it resulted in the discovery of an unconformity separating Proterozoic and Cambrian sandstones. In the Georgina Basin it led to the revision of a developing stratigraphic scheme, and to the recognition of a longer break between the Proterozoic and Cambrian than had previously been suspected.

Methods, sampling and results

Specimen processing

Both cores and hand samples were collected. Strike directions were determined with magnetic and solar compasses, and dips were measured with a Brunton compass. Usually eight, but sometimes more, core specimens (1" diameter, 7/8" length) were drilled from each hand sample in the laboratory. Before thermal demagnetisation the specimens were freed from

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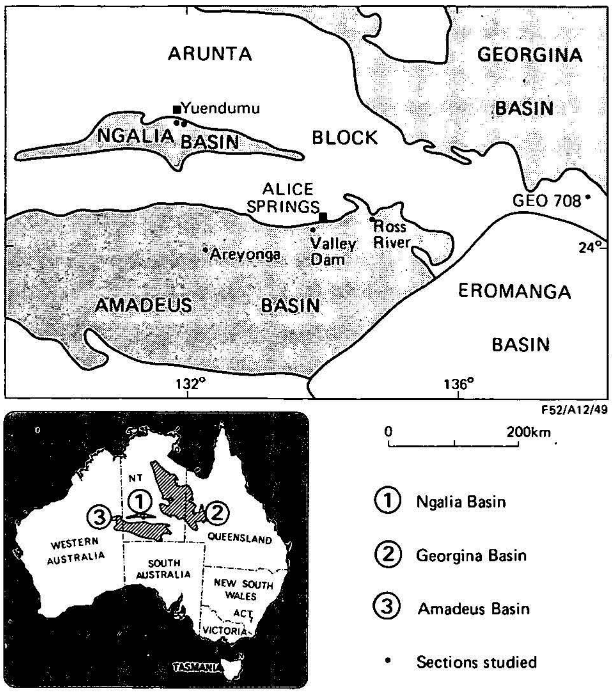


Figure 1. Locality map. Basin outlines are taken from the Tectonic Map of Australia and New Guinea, 1971.

air and connate H_2O in a desiccator-produced vacuum of 2.5-6.5 Pa. The cores were then refilled with argon gas in order to minimise chemical changes during the subsequent closely stepped (up to 13 steps) thermal demagnetisation treatment. Thermal demagnetisation of at least one core per hand sample also was carried out in argon gas environment (flow rate 0.5L/min.).

Heating was carried out in a magnetic field free space produced by 3-axis Rubens-Parry coils with feed-back control as described by McElhinny & others, 1971. Field control was usually + 5 nT and rarely exceeded + 10 nT during cooling.

To minimise viscous remanent magnetisation (VRM) effects samples were transferred in μ-metal containers to the field-controlled cryogenic magnetometer area, following the normal thermal demagnetisation procedures in the Black Mountain palaeomagnetic laboratory (ANU-BMR/Canberra).

Ngalia Basin

The geoolgy of the Ngalia Basin is described by Wells & others (1968, 1972), and Wells (1976). The unit studied in this basin is the lower part of the Yuendumu Sandstone, 7 km southeast of the Yuendumu Aboriginal Settlement (Figs 1, 2), on the Mt Doreen 1:250 000 Sheet area. About 700 m of the formation was measured in the type section Ex-7 (Wells & Moss, work in progress); the section is faulted but this is considered to be close to the true thickness. The lower unit is violet-brown sandstone and arkose, which rests unconformably on the Vaughan Springs Quartzite. The lower Yuendumu Sandstone was sampled along two sections 0.5 km apart (Section 1, Lat. 22°18′00″S, Long. 131°49′50″E; Section 2, Lat. 22°18′02″S, Long. 131°50′13″E).

In Section 1 thirty orientated hand samples were collected (ANTY1-ANTY30). Sample spacing was usually 2-3 m, rarely exceeding 4-5 m. Sampling in this section commenced directly above the unconformity at the top of the Vaughan Springs Quartzite, and covered the lower 100 m of the Yuendumu Sandstone.

Section 2 is located 0.5 km further east on the continuation of the same ridge of Yuendumu Sandstone across a small north-trending, fault-controlled valley. Nineteen oriented hand samples (ANYE1-ANYE19) were collected; sample spacing was as in Section 1. Outcrop conditions dictated that the sampling base of Section 2 had to be 50-60 m above the unconformity over the Vaughan Springs Quartzite. The top of Section 2 is stratigraphically higher than that of Section 1. The sampling overlap and short distance between the sections was designed to test the consistency of magnetic field reversals.

Rock-magnetism of the lower Yuendumu Sandstone.

The intensities of initial natural remanent magnetisation (NRM) vary between 268 mA.m^{-1} and 10 mA.m^{-1}, but usually cluster about 20 mA.m^{-1}.

Samples from the first few metres above the Vaughan Springs/Yuendumu Sandstone unconformity are very weakly magnetised. These samples show extreme directional changes during thermal demagnetisation, indicating the presence of two to three major vector-components commonly with random or almost opposite polarity, which are gradually broken down during thermal treatment. Since these extremely complex changes occur only within the first three metres above the unconformity they are attributed to chemical remanent magnetisation (CRM) components, and are probably associated with relicts of past groundwater circulations. Samples ANTY 1-3 produce normal groupings close to and above the magnetite Curie temperature; thus their initial magnetisation was probably normal.

Most samples from both Yuendumu Sandstone sections react to thermal demagnetisation treatment in a simple and consistent way. Initial intensities drop by about 50 percent after two-step heating (104°, 288°C), removing a probably relatively recent component (VRM + CRM). It is important to note that in some examples normal initial inclinations are steeper (at about 80°) than that of the present field. These magnetic changes cannot solely be attributed to recent VRM. They reflect a chemically grown (CRM) component, probably in iron hydroxides acquired during early Tertiary periods of deeply penetrating, intensive weathering (Burek, 1969, 1971; McElhinny & others, 1974; Schmidt & Embleton, 1976; Schmidt & others, 1977; Idnurm & Senior, 1978), which probably coincide with periods of tectonic stability. The possibility of subsequent tectonic tilt affecting the area of the Yuendumu escarpments should not be excluded. Slight southward tilt of Tertiary CRM-components would produce steeper inclinations (Burek & others, work in progress).

Subsequent changes of directions within individual samples on further heating (422°-655°C) are less significant, and are characterised by a gradual breaking down in intensity of a usually NNE-orientated, upward-directed inclination in the case of normal polarity; or a SSW-orientated, downward-directed inclination for reverse polarities. Stable directions beyond 422°C are interpreted as the original depositional remanent magnetisation (DRM). There is, however, considerable scatter in directions between different samples, probably caused by unfavourable conditions for alignment of the magnetic heavy minerals in the earth's field during deposition. Because of the DRM scatter no tilt or bedding corrections were applied. Use only of

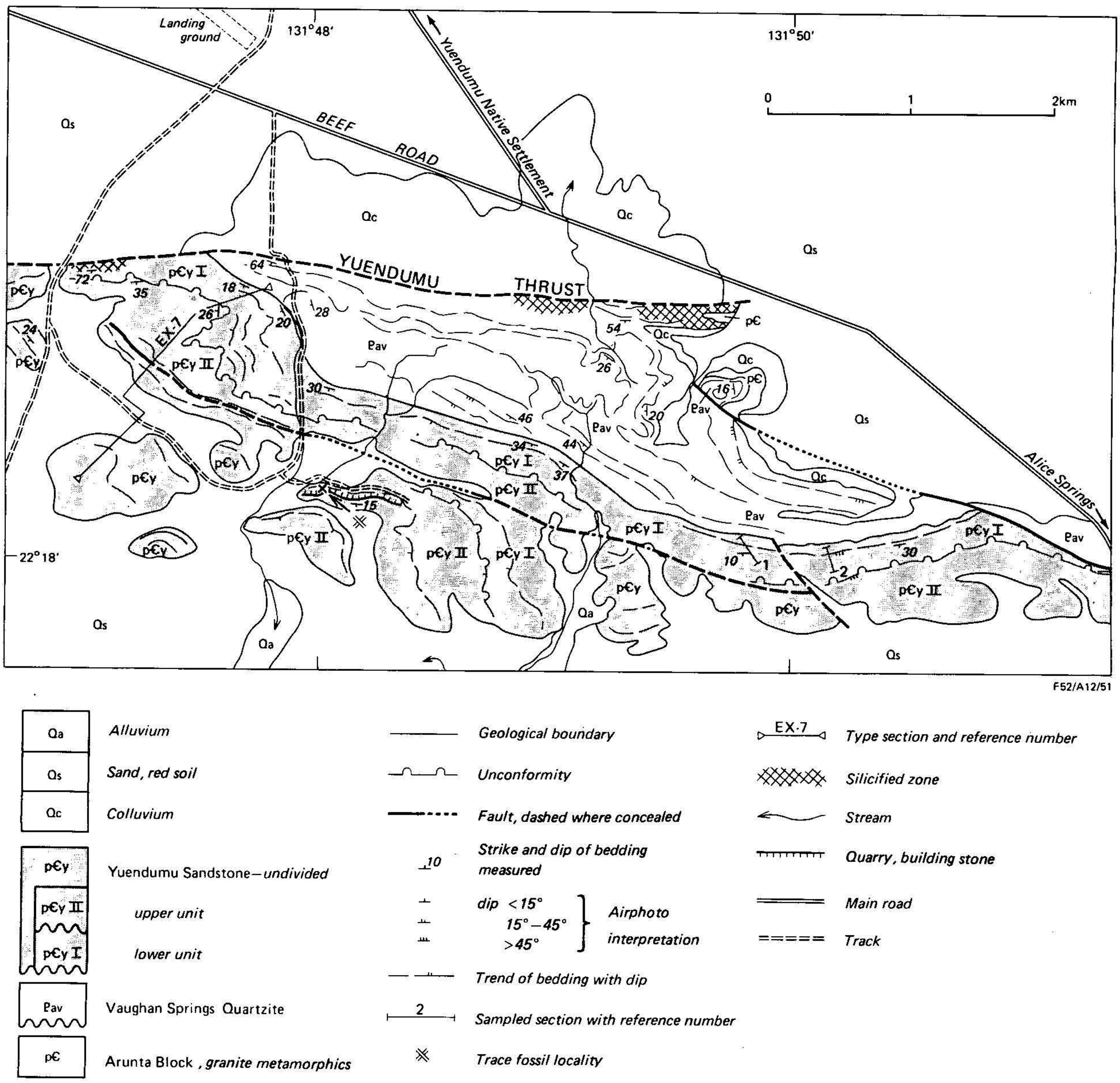


Figure 2. Geological map of the northern margin of Ngalia Basin south of Yuendumu Settlement, showing the location of two palaeomagnetic sampling sections and an unconformity recognised during this study. Geology modified from Wells (1972).

field-corrected NRM directions seemed preferable, because they allowed better interpretation of more recent CRM and VRM components.

A marked decrease in intensity of DRM occurs after the magnetite Curie temperature (570°C) transition. Directions above 570°C are usually consistent with previous ones (<570°C) up to temperatures of about 620°C. This reflects an original presence of hematite and magnetite in the heavy mineral fraction.

Magnetostratigraphy of the lower Yuendumu Sandstone. Vector-analysis of the Yuendumu Sandstone basically indicates the presence of two components:

The first is probably quite recent in origin, with a low blocking temperature, but it is not a VRM due to the present earth's field (inclination (i) = 55, declination (D) = NNE or SSW), since initial inclinations are too steep. It is thought to be primarily caused by intensive Tertiary, deeply penetrative weathering (Idnurm & Senior, in press) and probably also reflects a tectonic tilt component. This Cainozoic chemically grown and/or tilt component is destroyed after heating to 288°C.

The stable component of the second is similar to the first in declination, but inclinations are usually shallower. It is thought to be original Adelaidean DRM. Within individual samples this component is gradually and consistently broken down during thermal demagnetisation above the magnetite Curie temperature. It reflects two polarities. Since original DRM scatter in directions is large, it is difficult to produce a reliable Adelaidean palaeo-pole position (D = NNE or SSW, i = shallow).

In Section 1 (Fig. 3) the lowermost 2-3 m (ANTY 2-3) contain a complex multicomponent NRM (several CRM components), but appear originally to have been normally magnetised. The following 50-60 m (ANTY 4-16) are clearly and consistently normally magnetised (D = northerly orientated, i = up).

The first reversal (R1 = ANTY 17) (D = southerly orientated, i = initially steeply up, but progressively

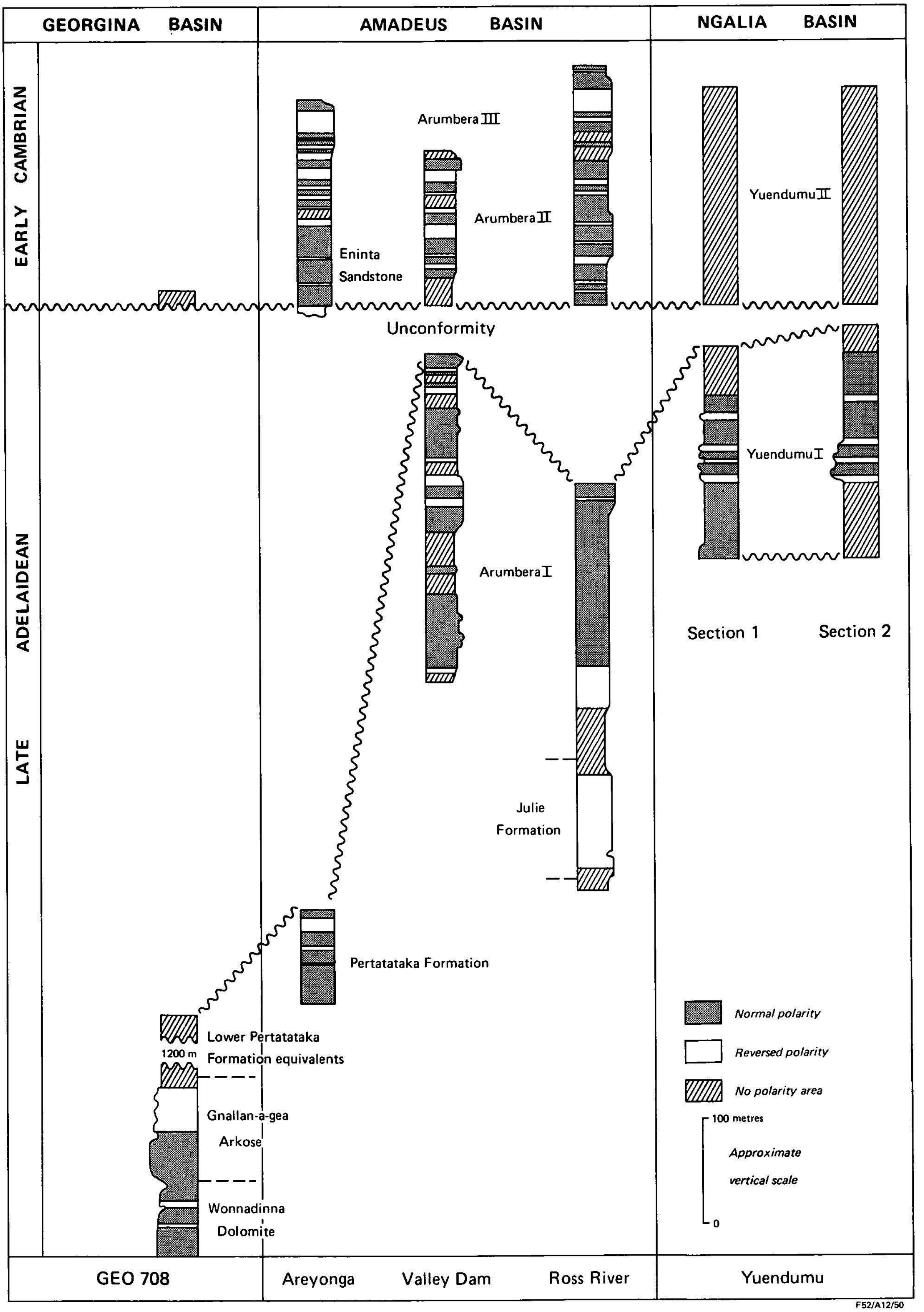


Figure 3. Magnetic polarity sequences from the Amadeus, Ngalia and Georgina Basins. The Amadeus Basin data are from Kirschvink (1978).

shallower during thermal demagnetisation) appears at about 60 m in Section 1, close to the crest of the Yuendumu Sandstone cliff. The next sample (ANTY 18) does not fit well into the general grouping of normal NRM's, and in addition does not represent an ideal case of reversed NRM directions (D = westerly orientated, i = downward); however, as it occurs next to a reversed sample (ANTY 17), and because of its clearly downward-directed inclination, we interpret the declination offset as being due to original DRM scatter and generally reflecting a reversed magnetisation of the earth's field during sedimentation. Sample ANTY 19 is normally magnetised. Thereafter three more reversals occur at 5 to 10 m intervals. Hand samples ANTY 20 (= R_2), 22, 23 (= R_3), possibly 27, and 29 (= R_4) contain stable, reversed DRM directions. Specimens from samples between the reversals are normally magnetised (Fig. 3).

Since all discussed reversed NRM directions are stability tested by thermal demagnetisation (626°C), they are interpreted as being original DRM components. Lightning and other VRM components are unlikely to survive demagnetisation treatment.

Sample ANYE 01 from Section 2 contains westerly orientated declinations and downward inclinations. It is an imperfect example of reversed DRM grouping, but is interpreted in the same way as ANTY 18 of Section 1 (= R_1). Sample ANYE 03 was taken about 2-3 m above the previously discussed sample and falls much better into the reversed DRM group (= R_2), as do samples ANYE 08.11 and 08.13 (= R_3), 11 (= R_4) and to a lesser extent 14 (R_5?). The higher parts of Section 2 appear to be normally magnetised. It appears that all four reversal horizons of the higher parts of Section 1 are present in the lower parts of Section 2 (Fig. 3). Section 2 is thus characterised by mixed polarity in its lower parts and normal magnetisation above.

In summary, at least four horizons of reversed polarity are persent in the middle parts of the lower Yuendumu Sandstone, but the lower Yuendumu Sandstone is predominantly normally polarised.

Georgina Basin

The late Proterozoic stratigraphy of the southern Georgina Basin will be described by Walter, and is summarised on Figure 5.

In this study, section GE0708 (Fig. 4) of the upper Wonnadinna Dolomite and the lower Gnallan-a-gea Arkose was sampled (Lat. 23°04′42″, Long. 137°33′36″). This section is 3.8 km ENE of Marqua Desert Bore, on the Hay River 1:250 000 Sheet area (Fig. 1).

The main dolomite unit (40 m) and the lower part of the overlying sandy and dolomite interval (20 m) were cored (ANTG1-ANTG38). Coring intervals were usually one metre, but in places were up to 3 metres. The higher parts of the siliciclastic sequence at the top of the Wonnadinna Dolomite were sampled by hand (ANGH1-ANGH6). Eight metres of section between the Wonnadinna Dolomite and the Gnallan-a-gea Arkose is not exposed.

The Gnallan-a-gea Arkose also was sampled by hand (ANGH7-ANGH14). Since the arkose (80 m) consists of at least four to five sedimentation cycles (conglomerate and fining-upward arkose sequences) and sedimentation rates were probably high, sample spacing for this reconnaissance survey was at about 10 m intervals. Conglomeratic layers were avoided, as they are unlikely to have acquired reliable DRM alignments. Sandstone layers were selected for sampling from the lowermost conglomerate cycles.

Rock-magnetism of the upper Wonnadinna Dolomite. The intensities of initial NRM in the first 40 m of the sampled section vary between 10 mA.m^{-1} and 0.2 mA.m^{-1}, but usually cluster around 2 mA.m^{-1}. The more intensively magnetised samples drop to less than 10 percent of the initial value during thermal demagnetisation to temperatures of 222°C. This very marked decrease in intensity is accompanied by pronounced and consistent directional changes—initially northerly orientated very steep inclinations (i = usually about 80°) become shallow at about 222°C. These initial NRM components are attributed to iron hydroxides, concentrated along cracks and joints in the dolomites (CRM).

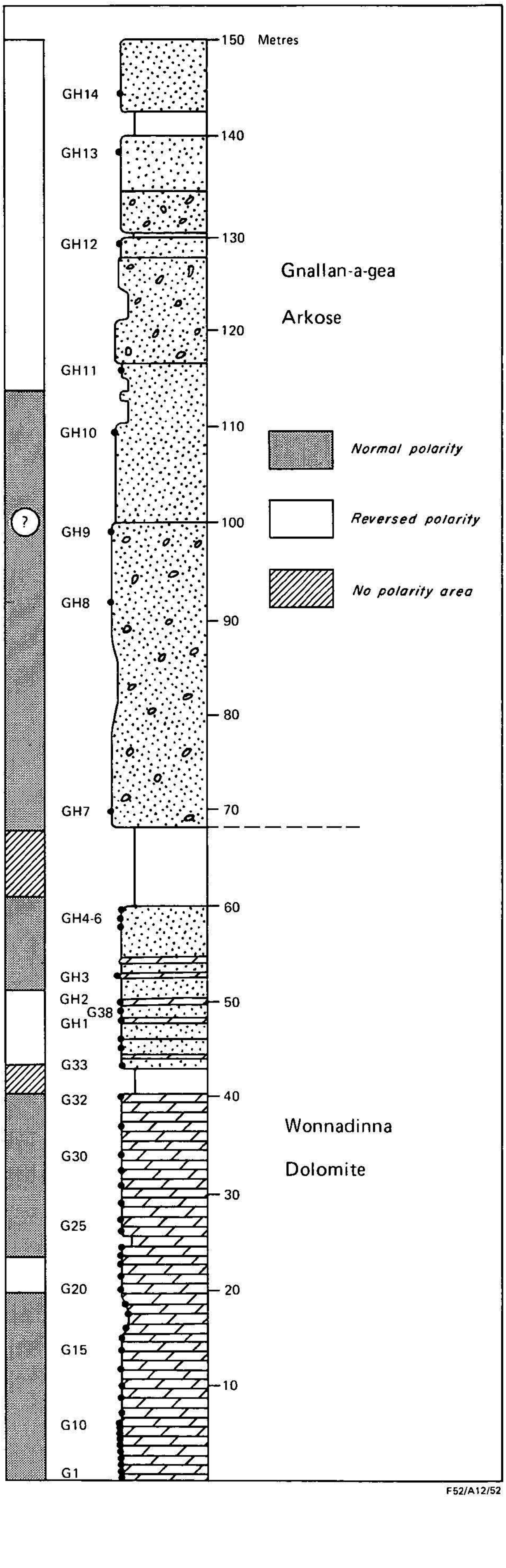
Subsequent thermal demagnetisation steps (255°, 288°, 300°, 333°C) show the gradual breaking down of generally NNW-orientated declination components, with shallow inclinations. Above 333°C most of the dolomite specimens exploded, in spite of repeated vacuum and argon-gas treatment, preventing further demagnetisation.

The uppermost Wonnadinna Dolomite consists of interbedded arkose, siltstone and dolomite (Fig. 4). Samples from these layers (cores: ANTG33-38, handsamples: ANGH1-6) are more strongly magnetised and can be demagnetised to hematite Curie temperatures. They indicate a basically similar two-component magnetisation to that described above, but the initially steep inclinations are very much less pronounced. Directional changes above 222°C are characterised by the gradual destruction of a consistently NNW-SSE orientated declination component with a shallow inclination; this component is considered to be the original DRM. These clastic and chemical rocks were deposited in a quiet lacustrine or marine environment, ideal for palaeomagnetic purposes. This suitable environment is the probable cause for the good DRM groupings of the two opposite polarities (D = NNW-SSE, i = shallow). Because the stable and completely demagnetised DRM components of the clastic Wonnadinna sequence are similar to the more stable component in the only partially demagnetised dolomite sequence it is concluded that these also reflect original DRM components.

In summary, the dolomite is predominantly normally magnetised. Only specimens ANTG20.11 and 21.11 produce southwesterly orientated, i.e. reversed, directions with shallow inclinations. The basal parts of the siliciclastic Wonnadinna sequence are reversely magnetised (D = SSW, i = shallow), the higher parts have normally orientated directions (D = NNW, i = shallow). No attempt has been made to calculate palaeo-pole positions because the results of chemical and AC-demagnetisation measurements have not yet been fully analysed, and—with the exception of the Wonnadinna siliciclastics—DRM scatter is large.

Rock-magnetism of the lower Gnallan-a-gea Arkose. The intensities of initial NRM vary between 35 mA.m^{-1} and 0.8 mA.m^{-1}, but usually cluster at about 10 mA.m^{-1}.

In comparison with the Wonnadinna Dolomite, the decay of intensity at temperatures less than 222°C is much less pronounced. This is comparable to the demagnetisation pattern of the Wonnadinna siliciclastics. Also, the drastic initial directional changes—reflecting a recent CRM component in the Wonnadinna



Dolomite and the Yuendumu Sandstone of the Ngalia Basin—are present in only a few samples of the Gnallan-a-gea Arkose.

A pronounced drop in intensity is observed at the magnetite Curie temperature transition. Directions below and above the magnetite Curie temperature are similar or identical. In all probability this reflects the presence of hematite and magnetite during deposition. Directional changes during thermal demagnetisation are usually straightforward—NNW (D) for normal polarities, and SSE (D) orientated vectors for reversed polarities, both with shallow inclinations, are gradually broken down during treatment. These directions are considered to reflect the original DRM. The DRM scatter is attributed to rapid deposition of coarse sediments, which is unsuitable for good DRM alignments. The Gnallan-a-gea DRM groupings are similar to those observed in the Wonnadinna Dolomite.

The lowermost part of the Gnallan-a-gea Arkose is normally magnetised. There is a possibility of a reversal at the top of the first coarse conglomerate unit. Specimen ANGH 09.22 only poorly reflects reversed DRM directions, and could be interpreted as a case of extremely poor DRM alignment during sedimentation in a normally polarised Earth's field.

The upper parts of the Gnallan-a-gea Arkose in Section GEO708 appear to be reversely magnetised.

Magnetostratigraphy of Section GEO708. Vector analyses of the Wonnadinna Dolomite and the Gnallan-a-gea Arkose indicate:

1. That the dolomite beds contain two components. The first, characterised by very steep inclinations, is probably quite recent in origin. But it cannot be a VRM caused by the present earth's field, since inclinations are much too steep. It is thought to be caused primarily by intensive Tertiary, deeply penetrative weathering (Idnurm & Senior, in press) and also to reflect a tectonic tilt component. This Cainozoic chemical and tilt component is destroyed after heating to 222°C. The second has stable NNW-orientated DRM components, with shallow inclinations. They reflect two polarities. Because the DRM scatter is broad and impossible to completely demagnetise thermally, it is difficult to produce a reliable palaeo-pole position.
2. DRM groupings of two polarities from the siliciclastic Wonnadinna sequence are much smaller and similar to those of the dolomite. The initially well preserved very steep inclinations in the dolomite are much less pronounced in the clastics.
3. The Gnallan-a-gea Arkose basically reflects one vector only, thought to be original DRM. DRM alignments are again similar to the Wonnadinna sequence (D = NNW-SSE, i = shallow), but DRM scatter is broad. Within individual samples this vector is gradually and consistently broken down during thermal demagnetisation above the magnetite Curie temperature. Two polarities are present in the sampled sediment sequence.

The upper Wonnadinna Dolomite is predominantly normally magnetised (D = NNW), but has a reversal in its middle part. The siliciclastic unit at the top of the Wonnadinna Dolomite is reversely magnetised at its

Figure 4. Section GEO708 from the southern margin of the Georgina basin showing thicknesses in metres (on the right) and palaeomagnetic sampling sites and the interpreted polarity spectrum (on the left).

base (D = SSE), and normally magnetised at its top (D = NNW). The lowermost Gnallan-a-gea Arkose seems to be predominantly normally magnetised. The higher parts of the sampled section are consistently reversely magnetised.

It is of interest to consider the possibility of a hiatus between the palaeoenvironmentally dissimilar Wonnadinna Dolomite and Gnallan-a-gea Arkose. Since both units basically produce two polarities with quite similar DRM directions, i.e. reflect similar magnetic palaeo-pole positions, there are no palaeomagnetic indications of a major time-break between the two sequences.

Magnetostratigraphic correlations and new lithostratigraphic observations

Kirschvink (1978) determined a latest Adelaidean to earliest Cambrian magnetic polarity sequence through sections in the Ross River, Valley Dam and Areyonga areas of the Amadeus Basin (Fig. 1). His results are depicted in Figure 3. In the Ross River area the Julie Formation (dolomite) and the basal part of the Arumbera Sandstone I are predominantly of reversed polarity. The remainder of Arumbera Sandstone I is a longer interval of predominantly normal polarity. The normal interval is also present in the Arumbera Sandstone I of the Valley Dam area, but there it is followed by a sequence of mixed polarity (still in Arumbera Sandstone I). In the Ross River area, deposits of this mixed interval were either eroded off or were never deposited. A widespread unconformity (possibly due to the Petermann Ranges Orogeny) occurs near the base of the Cambrian in the Amadeus Basin (Wells & others, 1970; Daily, 1972). Above the unconformity the magnetic polarity sequence consists of mixed normal and reversed zones, with an initial bias toward normal polarity gradually disappearing upwards. Arumbera Sandstone II and III are siltstone and sandstone with abundant trace fossils.

At present it is not possible to use brief polarity events for correlation. Only when there is some independent control can correlations within mixed polarity intervals be attempted on a regional scale (Burek, 1964, 1967, 1968). However, long-lasting polarity events are more easily recognised and distinguished from each other (Burek, 1970; McElhinny & Burek, 1971). It is these coarse polarity sequences that can be used in comparing the history of the Amadeus, Ngalia and Georgina Basins.

Ngalia Basin

The Yuendumu Sandstone is overlain, probably unconformably, by the Lower Cambrian Walbiri Dolomite, and itself unconformably overlies the Vaughan Springs Quartzite (Fig. 5). Regional stratigraphic relationships suggest that the Yuendumu Sandstone is younger than the upper of the two Proterozoic tillites (see Preiss & others, 1978). Abundant trace fossils collected from Yuendumu Sandstone II in the Yuendumu Quarry by Wells have been identified by Walter as *Planolites ballandus* Webby, a form also found in the Arumbera Sandstone III of the Amadeus Basin. On the basis of lithology and stratigraphic position, supported by the occurrence of trace fossils, the Yuendumu Sandstone has been correlated with the Arumbera Sandstone.

The polarity pattern of the lower Yuendumu Sandstone can be used to exclude one possible correlation: the time interval characterised by long-lasting, predominantly reversed polarity reflected in the Julie Formation and the lower part of Arumbera Sandstone I is not represented in the Yuendumu Sandstone.

Correlations are possible with: the upper Arumbera Sandstone I; the Arumbera Sandstone II; the Arumbera Sandstone III; and the Eninta Sandstone (equivalent to Arumbera II and III). In other words, using the magnetostratigraphic evidence only, it is not possible to decide whether the lower Yuendumu Sandstone is latest Adelaidean or Early Cambrian in age.

Because an unconformity separates Arumbera Sandstone I from Arumbera Sandstone II, we decided to look for a similar break within the Yuendumu Sandstone, in an attempt to resolve the ambiguities in correlation. An airphoto interpretation of the critical outcrops of the Yuendumu Sandstone was undertaken by C. J. Simpson (BMR). An unconformity with slight angular discordance was discovered between the upper Yuendumu Sandstone with trace fossils, and the lower Yuendumu Sandstone with a mixed polarity pattern (Fig. 2). This interpretation was then confirmed in the field by A. T. Wells. At the unconformity there is an abrupt change of rock type. The top of the lower unit is dark red-brown medium-grained sandstone and granule conglomerate; the base of the upper unit is coarse grained, yellow-weathering, felspathic, tough, siliceous sandstone with abundant large clay pellets and some discontinuous conglomerate beds. The unconformity surface is uneven.

Using the unconformity, the trace fossils, and the palaeomagnetic data, we therefore correlate the lower Yuendumu Sandstone with the upper part of Arumbera Sandstone I, and Yuendumu Sandstone II with Arumbera Sandstone II/III. Thus Yuendumu Sandstone I is late Adelaidean and Yuendumu Sandstone II is Early Cambrian in age.

Georgina Basin

The Wonnadinna Dolomite–Gnallan-a-gea Arkose couple is comparable to and prior to this work had been correlated by Walter with the Julie Formation–Arumbera Sandstone I couple. Constraints are provided by the correlation of the Yardida Tillite with the tillites of the Areyonga Formation (Preiss & others, 1978) and the discovery of Early Cambrian fossils in a younger unit (Fig. 5).

As can be seen in Figures 3 and 4, the magnetostratigraphics results strongly suggest that it is not possible to correlate the Wonnadinna Dolomite with the Julie Formation, and the Gnallan-a-gea Arkose with Arumbera Sandstone I. It can be suggested that the units are diachronous but still lithologically correlative, but subsequent studies resulting from the magnetostratigraphic work indicate that such is not the case (see below).

The age of the Grant Bluff Formation, which conformably overlies the Gnallan-a-gea Arkose in the southern and southwestern Georgina Basin, has long been problematic. It had been supposed to be Early Cambrian, but no convincing or distinctive fossils have been found. The evidence for an Early Cambrian age was an occurrence of fossils in the Barrow Creek 1:250 000 Sheet area (Smith, 1972, p. 68) in a unit now known to be separate from and younger than the Grant Bluff Formation (Walter, work in progress).

The magnetostratigraphic results indicated that the Grant Bluff Formation and its underlying units may be older than had been considered (a younger age is not possible, because of the established Early Cambrian age

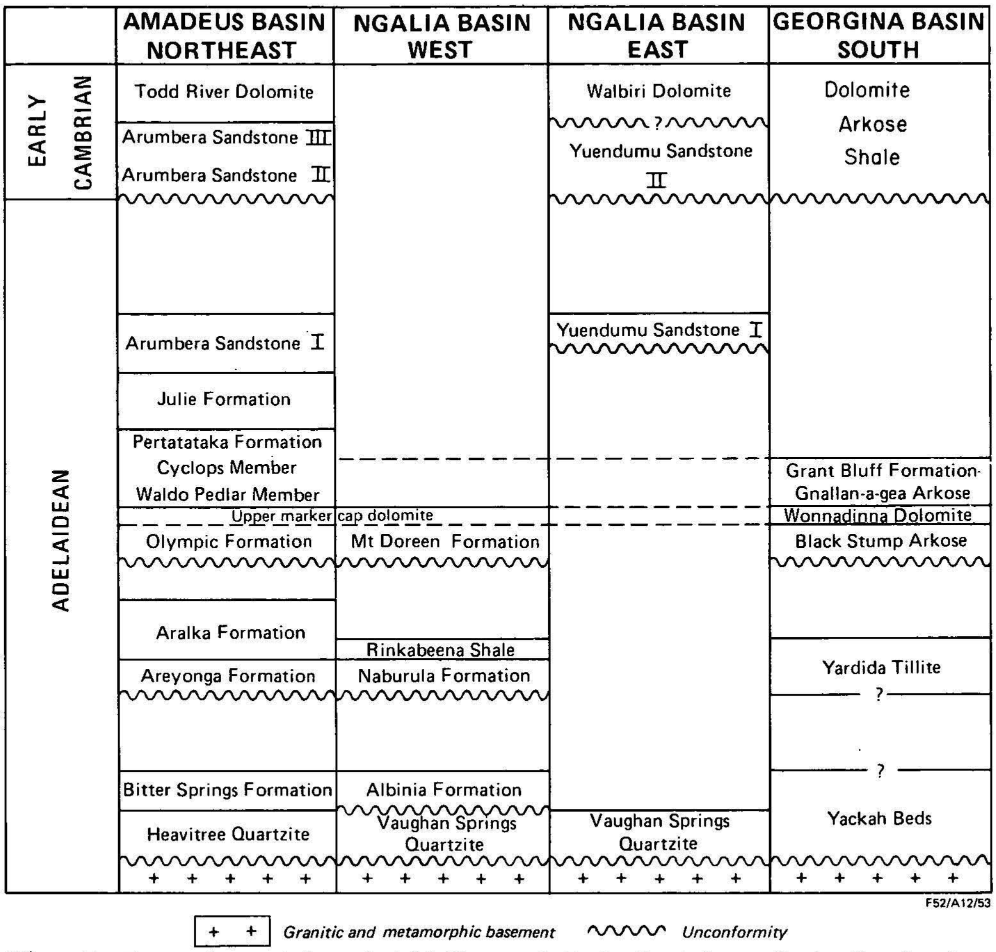


Figure 5. Proposed correlation of Adelaidean and Early Cambrian units in the Amadeus, Ngalia and Georgina Basins.

of an overlying unit—see Figure 5). This led to reconsideration of the observation by Shaw & others (work in progress) that a unit resembling the Grant Bluff Formation occurs at depth in two drill holes near Mount Skinner on the Alcoota Sheet area. These holes spudded in at a stratigraphic level near that containing the Mount Skinner fauna of Wade (1969). This same fauna occurs in Arumbera Sandstone I. Re-examination by Walter of Centamin cores DDC1 and DDC2 from 1.8 km and 8.2 km east of Mount Skinner (Alcoota 1:250 000 Geological Series Sheet) confirmed that the Grant Bluff Formation occurs at depth, below equivalents of Arumbera Sandstone I, the Julie Formation and the upper Pertatataka Formation. It is apparent that the Grant Bluff Formation is a correlative of the lithologically very similar Cyclops Member (Pertatataka Formation) of the Amadeus Basin (Fig. 5).

Thus it seems that the Gnallan-a-gea Arkose is a correlative of the Waldo Pedlar Member (Pertatataka Formation), and the Wonnadinna Dolomite correlates with the dolomites of the upper Olympic Formation (upper tillite) of the Amadeus Basin (Fig. 5). A consequence of these correlations is that in the Hay River 1:250 000 Sheet area of the southern Georgina Basin there are no equivalents of the upper Pertatataka Formation, the Julie Formation or the lower Arumbera Sandstone, and there is a major break between the Adelaidean and Cambrian, of longer duration than elsewhere in central Australia.

A further consequence of these correlations is the equating of the Wonnadinna Dolomite with the dolomite capping the tillite of the Mount Doreen Formation (upper tillite) in the Ngalia Basin (see Preiss & others, 1978). There are some preliminary palaeomagnetic data from the Mount Doreen dolomite in the Patmungala Syncline which indicate SW and NE declinations and extremely shallow inclinations for the DRM (Burek & others, work in progress). These declinations are different from those of the upper Wonnadinna Dolomite, but this difference may have been caused by clockwise rotation of the Patmungala Syncline area during the Carboniferous Mount Eclipse Orogeny (Wells & Moss, work in progress). In addition, it may be that the Mount Doreen Formation dolomite correlates with dolomites lower in the Wonnadinna sequence which have not yet been studied palaeomagnetically.

Conclusions

The magnetostratigraphic study has allowed a refinement of our knowledge of the Adelaidean to Early Cambrian stratigraphy of the Ngalia and Georgina Basins. In addition it has extended back in time the polarity sequence of Kirschvink (1978), and provided palaeomagnetic data of relevance in correlating the upper glacial sequence in the Adelaidean of central Australia. DRM directions from the sequences studied are in agreement with the established Adelaidean to Cambrian apparent polar wander path (McWilliams, 1977), but since the DRM scatter is large our data cannot be used to refine the known path.

There is an unconformity within the Yuendumu Sandstone of the Ngalia Basin. The lower part of the Yuendumu sandstone (I) can be correlated with the upper part of Arumbera Sandstone I of the Amadeus Basin, and is latest Adelaidean in age. The upper part, containing trace fossils, correlates with the Early Cambrian part of the Arumbera Sandstone (probably part III).

In the southern Georgina Basin, in the Hay River sheet area, there is a break of long duration between Early Cambrian units and the Grant Bluff Formation of Adelaidean age. Figure 5 shows our interpretation of the Adelaidean to Early Cambrian stratigraphy of the Ngalia and Georgina Basins.

Acknowledgements

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Mode of emplacement of the Papuan Ultramafic Belt

J. B. Connelly

Geological and geophysical evidence from a number of studies has established that the Papuan Ultramafic Belt is probably an overthrust sheet of oceanic crust and mantle, with a thicker crustal section than normal oceanic crust. Earlier workers have described the overthrust as the result of north-south compression produced by the northwards movement of the Australian Plate. However, left-lateral faulting of the Belt subsequent to emplacement is evident from the displacement of the main ultramafic bodies. Reconstruction of the Belt by reversing the movements along these faults suggests that the Belt was aligned north-south when originally emplaced. Theoretical cross-sections of the Belt along three profiles were contructed—assuming that is was originally aligned north-south with a shallow easterly dip, and was sheared in a northwesterly direction. The computed gravity and magnetic anomalies along these theoretical cross-sections match the observed anomalies closely. The Belt was probably emplacd 30° south of its present latitude, before the Australian Plate started to move north about 55 m.y. B.P.

The thick Cretaceous crust which forms the Belt is thought to extend north under most of the western part of the Solomon Sea, and east along the Woodlark Rise where is has been subject to extensive rifting. The presently exposed part of the Belt is estimated to have been uplifted by up to 10 km.

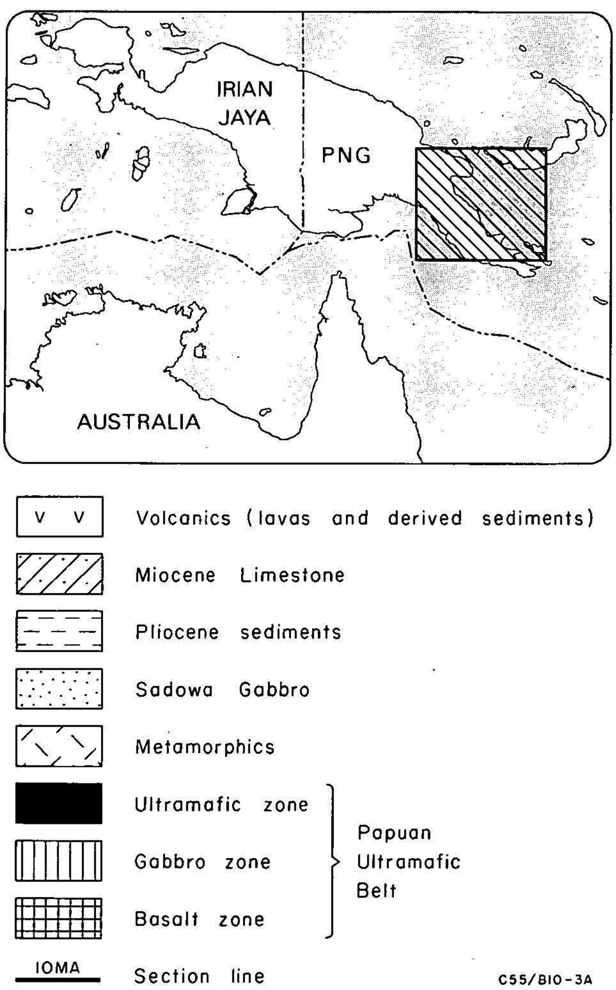
Introduction

The Papuan Ultramafic Belt (Fig. 1) is one of the best preserved peridotite-gabbro-basalt complexes in the world. Thompson (Thompson & Fisher, 1965) first suggested that the Belt was an overthrust sheet of oceanic crust and mantle, and this hypothesis was supported by later workers (Davies, 1968, 1971; St John, 1967; Milsom, 1971). The Belt crops out over a distance of some 400 km along the northeastern side of the Papuan Peninsula (Fig. 1), and is elevated in parts to 2500 metres above sea level.

The detailed geology of the Belt has been described by Davies (1971), and a description of the geology of the whole Papuan Peninsula is given by Davies & Smith (1970). The Belt is composed of three zones: an ultramafic layer, overlain by a gabbroic layer, which is in turn overlain by a basaltic layer. The ultramafic layer is mainly made up of harzburgite, with some dunite and enstatite-pyroxenite. Pods of material in which there is evidence of crystal settling (cumulus phases) are present at the top of the layer; at some locations these grade upwards into cumulus phases of the gabbroic layer.

The gabbroic layer is a complex assembly of granular, microgranular and cumulus phases. The granular and microgranular phases intrude each other and the granular phases also intrude the cumulus phase. Cumulus rocks may make up as much as 30 percent by volume of the gabbroic layer; there is evidence that the abundance of plagioclase in cumulus phases increases towards the top of this layer. The basaltic layer is predominantly massive basalt, with some submarine lavas; its contact with the gabbro is sometimes distinct and sometimes gradational. The contact is frequently intruded by tonalites (not shown on Fig. 1), which are younger than the Belt.

The age of formation of the complex is regarded as Jurassic and/or Cretaceous (Davies & Smith, 1970). Potassium-argon ages of 116 m.y. have been obtained from pyroxenes in the basaltic layer, and two others of 147-150 m.y. from the gabbros. Late Cretaceous foraminifera have been found in pockets of marl associated with the basaltic layer. Potassium-argon ages of 50-55 m.y. for the tonalites which intrude the basaltic-gabbro contact (Davies & Smith, 1970) set an upper limit on the age of formation of the complex.



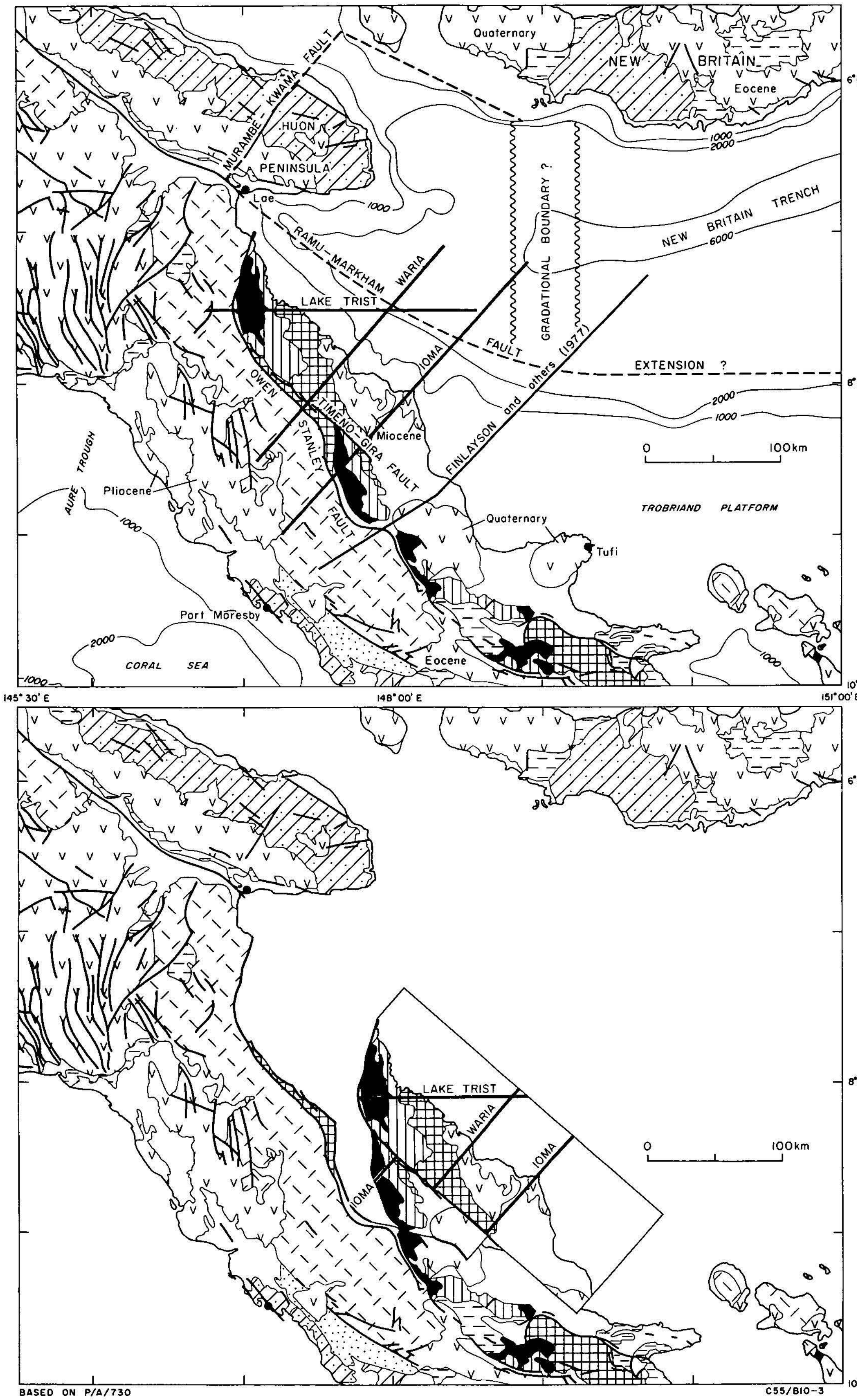


Figure 1(a). Geological and locality map of the East Papua Region, showing lateral extent of the thick Cretaceous crust. 1(b). Reconstruction of the Papuan Ultramafic Belt prior to shearing. For legend see p. 57.

Dacitic and basaltic lavas and derived sediments were laid down on the Belt during the Eocene; the environment in which these were laid down is not clear. A break in sedimentation then occurred before deposition in the Middle Miocene of shallow-water tuff and agglomerate.

To the south and west the Belt abuts the Owen-Stanley metamorphics; the boundary between the two is marked by the Owen Stanley Fault system. Pieters (1974) has subdivided the Owen-Stanley metamorphics into an eastern and a western group. The eastern group is considered to be mainly composed of metabasalts and is presumably derived from a part of the ocean floor which was continuous with the Papuan Ultramafic Belt. The western group is composed of pelitic and psammitic metasediments, which grade westwards into turbidite-type deposits of continental provenance. The general mineral assemblage suggests a high-pressure low-temperature environment, with greenschist facies predominating (Davies & Smith, 1970). Blue-schist facies rocks are present near the Owen Stanley Fault system.

Outcrops of rocks of the ultramafic belt and of the Owen-Stanley metamorphic occur on islands to the east of the Peninsula, but the structure of all the area east of 148°30′ is complex and is not well understood. Several large Quaternary volcanic centres are present along the northern margin of the Belt between it and the Trobriand Platform (Fig. 1a). Quaternary sedimentary cover is present along both the north and south coasts of the Papuan Peninsula, and on islands in the Trobriand Platform.

Gravity measurements have been made in the region (St John, 1967; Milsom, 1971); the average gravity station coverage is now about one station per 60 km^2, although the coverage is very variable. Figure 2 is a contoured map of the gravity field showing Bouguer anomalies over land, and free-air anomalies offshore. Aeromagnetic coverage of the whole of the Papuan Ultramafic Belt has been completed (CGG, 1969, 1971, 1973) (Fig. 3). These surveys were flown at heights of 4500 metres over the Papuan Peninsula and 2500 metres over the Trobriand Platform. Flight lines were mainly north-south with a separation between the lines of between 5 km and 15 km. A semi-detailed seismic refraction survey of the whole area was undertaken in 1973, and a structural interpretation of the data from this survey is given by Finlayson & others (1976, 1977).

The hypothesis that the Belt is an overthrust slab of oceanic crust and mantle is adopted in this paper. However, a different hypothesis for the mode of emplacement of the Belt is proposed. Davies (1971, 1977) implies that the overthrust is the result of NNE-SSW compression produced by northward movement of the Australian Plate during the Tertiary. In this paper east-west compression resulting from the westerly motion of the Pacific Plate is suggested as the cause of the overthrust. Overthrusting occurred before the Australian Plate started to move north when tectonic events along its northern margin were controlled entirely by the westerly motion of the Pacific Plate. The Belt was originally part of the Pacific Plate, and was overthrust from the west onto the edge of the Australian Plate.

Evidence is presented that the crust which forms the Papuan Ultramafic Belt also forms the floor of the western Solomon Sea and part of Huon Peninsula. It is suggested that the Ramu-Markham fault zone may con-

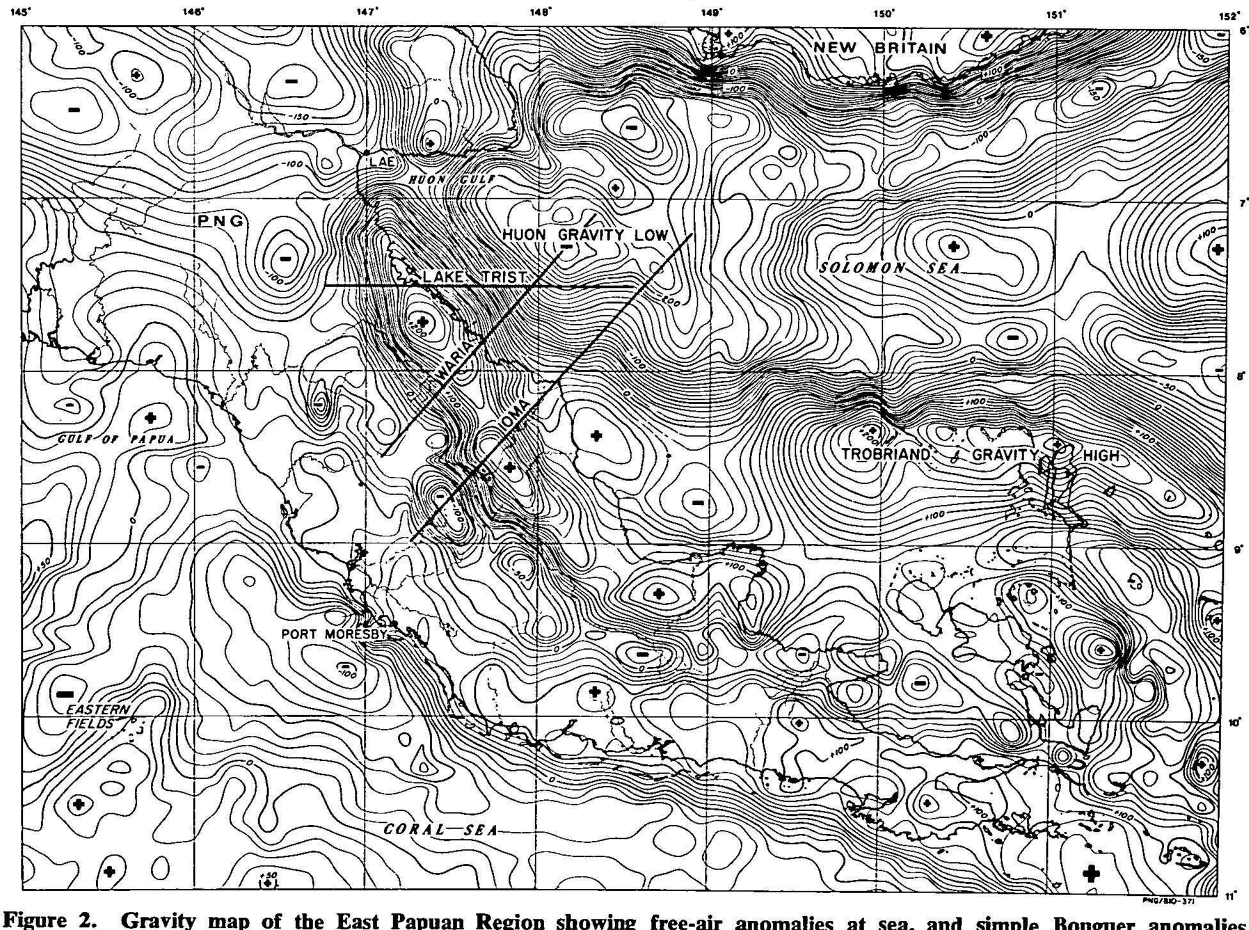


Figure 2. Gravity map of the East Papuan Region showing free-air anomalies at sea, and simple Bouguer anomalies (Bouguer density 2.67 tm^{-3}) on land.

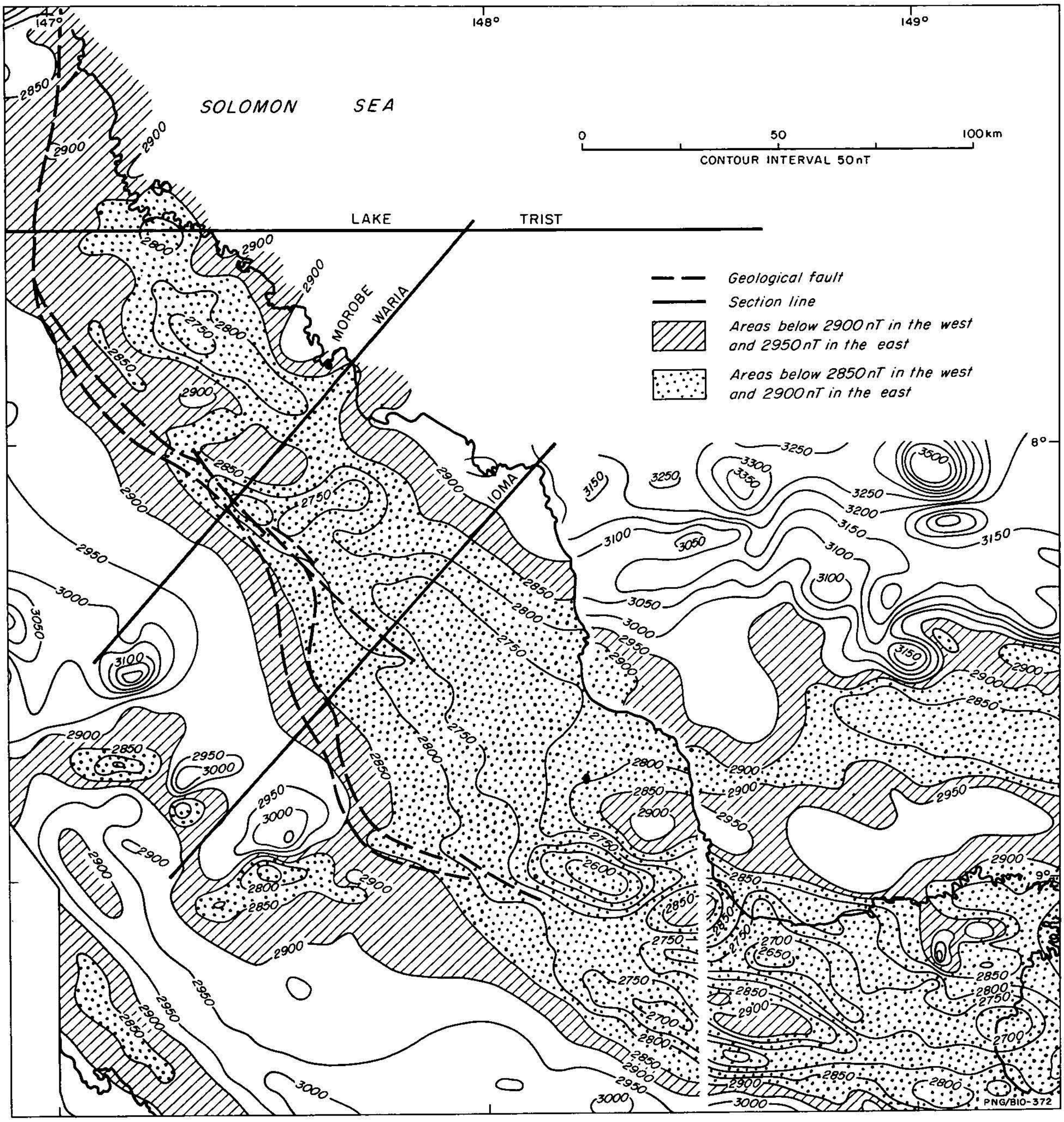


Figure 3. Aeromagnetic anomaly map of the Papuan Ultramafic Belt. The eastern part was flown at 2500 metres and the western part at 4500 metres.

tinue eastwards along the north coast of the Papuan Peninsula, and along the northern edge of the Trobriand Platform.

Emplacement from the west: evidence from geological reconstructions

Overthrusting of the Papuan Ultramafic Belt along sections of the Owen Stanley Fault (Fig. 1a) is indicated by the existence of ultramafic rocks presumably of mantle origin, overlying metasediments of the Owen Stanley Ranges. The Owen Stanley fault strikes in a general northwest-southeast direction, which has led to the assumption that the direction of thrusting must be northeast-southwest. However, in detail the fault is made up of alternating north to NNE, and northwest-southeast-striking sections. The north to NNE sections are those along which ultramafic rocks overlie metasediments (Fig. 1) and these represent the original thrust. The northwest-southeast sections are left-lateral faults, not thrusts; Davies (1971) has identified the Gira Fault—which is the longest northwest-southeast section (Fig. 1)—as a left-lateral fault with a displacement of 90 km, and it seems reasonable to assume that the other northeast-southwest sections also represent left-lateral faults. Reversal of the left-lateral movement along the Gira fault and along other northeast-southwest faults farther south leads to the reconstruction of the Papuan Ultramafic Belt shown in Figure 1b. The Belt was thus aligned north-south prior to the left-lateral shearing.

From this reconstruction a westward thrusting of the Belt appears as the most likely method of emplacement. Consequently both the emplacement of the Belt and the subsequent left-lateral shearing result from east-west compressional forces. This is in conflict with Davies' (1971) statement that 'a subsidiary effect (of the north-

	Measured samples Kroenke & others	*Nafe-Drake Curve*	*Birch relation mean atomic weight 21*	*This paper*
Basalt		2.63	2.48	2.60
Gabbro	2.95-2.98	2.84	2.84	2.85
Ultramafic	2.91-3.15	—	3.18	3.33

Table 1. Comparison of rock densities derived from various sources, in tm^{-3}.

Seismic velocities used were 5.66 km/s, 6.86 km/s and 7.96 km/s for the basalt, gabbro, and ultramafic layers respectively (Finlayson & others 1976), and the Birch relation used was $Y\rho = -2.55 + 3.31\ \rho$.

ward movement of the Australian Plate) may well have been the thrusting of a plate of oceanic mantle and crust (the Papuan Ultramafic Belt) southwards over the Cretaceous geosynclinal sediments of what is now eastern Papua'.

It can be argued that the reconstruction shown in Fig. 1b produces an arc concave to the northeast. However, that part of the Belt east of 148°30′ is very complex, and no clear thrust plane is apparent. Unmetamorphosed basalt of Cretaceous and Eocene age are present south of the Belt in this area (Fig. 1); these basalts probably represent the southern margin of the thrust plate which was not itself subjected to thrusting. This part of the Belt was elevated to its present position by later Middle Miocene uplift, which affected the whole region.

The arc-like appearance of the reconstruction could be reduced by including several smaller left lateral faults in the reconstruction. In addition the reconstruction could be extended to include a section of the thrust which is assumed to underlie the eastern part of the Huon Peninsula (see section on lateral extent of the thick Cretaceous crust); if this were done the northerly trend of the thrust plane would be accentuated.

Gravity and magnetic modelling methods

Two-dimensional magnetic and gravity models were constructed along the three sections shown on Figures 2 and 3. The gravity was modelled by adjusting the cross-section of the bodies manually until the observed and calculated profiles matched. The magnetics were modelled almost entirely by computer. An approximate starting model was adopted, which was then adjusted by successive iteration. Iteration continued until the standard deviation between the observed and calculated profiles ceased to change significantly. In practice considerable experimentation is required to find a starting model which gives a solution that both converges and is geologically reasonable. The program is based on a general non-linear least-squares program developed by G. Gibson (pers. comm.).

Table 1 shows the density values of rocks from the Papuan Ultramafic Belt, derived both from seismic velocities, using the relations of Birch (1961) and Nafe & Drake (1963); and from laboratory measurements on 12 rocks from two localities in the Papuan Ultramafic Belt (Kroenke & others, 1974). Models of the structure within the crust only, ignoring isostatic compensation of the masses involved, were produced using the densities: basalt 2.6 tm^{-3}, gabbro 2.85 tm^{-3} and ultramafic 3.33 tm^{-3}. Models which included compensating masses at the base of the crust were also constructed, and it was found that much greater densities were required for the basaltic and gabbroic layers in order to retain the same basic crustal structure. The densities used for these models were basalt 2.7 tm^{-3}, gabbro 2.95 tm^{-3} and ultramafic 3.33 tm^{-3}.

The density of 2.6 tm^{-3} for the basalt layer is approximately equal to that from the Nafe-Drake curve, but this density and that from the Birch relation are low, as all known constituents of basalt have densities of 2.6 tm^{-3} or greater. The density of 2.85 tm^{-3} for the gabbroic layer approximates both the Nafe-Drake and the Birch relations. This density is considerably less than that found from the laboratory measurements (Kroenke & others, 1974). A standard upper mantle density of 3.33 tm^{-3} was adopted for both the compensated and uncompensated models. However, the values measured both in the laboratory, and given by the Birch relation, were lower than this (Table 1). The larger value was adopted to fit the observed gravity along the northeast coast, where the layer interfaces are known from seismic information.

West of the Owen Stanley Fault a uniform model was adopted on all three sections, consisting of a 2.6 tm^{-3} layer extending to 6 km, and a 2.85 tm^{-3} layer extending to the Moho at 32 km. The standard crust adopted was that of Finlayson & Cull (1973).

For the magnetic models a magnitude of 2 A m^{-1} was adopted for the remanent vectors; angles of both −31° and −60° were used for the inclination. The magnitude is of the same order as measured values for samples taken from the basaltic layer (unpublished BMR data). The inclination of the vector was initially assumed to be the same as the present Earth's field (i.e. −31°) but, in the later stages of the interpretation, a value of −60° was used.

Emplacement from the west: evidence from gravity and magnetic modelling

The most northerly part of the Owen Stanley Fault is a north-south striking thrust plane. Gravity modelling along an east-west traverse across this area (Fig. 4, Lake Trist Section) indicates that the various contacts dip east at the following angles: thrust plane 23°, ultramafic gabbro 16°, gabbro-basalt 13°, basalt-sediments 12°. The model computed for the Lake Trist section was assumed to represent the approximate cross-section along the entire length of the original north-south thrust; theoretical cross-sections (Fig. 5) along two traverses perpendicular to the left-lateral faults were derived by geometrical calculation. Vertical discontinuities were included at the left-lateral faults. The top section (Fig. 5) is equivalent to the Trist Section. The middle section (Fig. 5) is the theoretical section along the Waria traverse and the bottom section (Fig. 5) is the theoretical section along the Ioma traverse.

Comparison of the theoretical cross-sections with those obtained by modelling the actual gravity anomaly (Fig. 4) shows that they are very similar. This provides further evidence that the Belt was originally emplaced north-south and was shearing in a northwest-southeast direction. The main discrepancy between the theoretical and computed models is on the Ioma section (Fig. 4) between the coast and the Gira Fault. The theoretical model predicts that the layers should be horizontal in this area, whereas the computed model shows that they dip away from the Gira Fault. The amount of dip on this part of the computed model indicates that about 10 km of uplift could have occurred in this area. The uplift probably occurred in the Miocene (Davies, 1971) after the thrusting and shearing of the Belt.

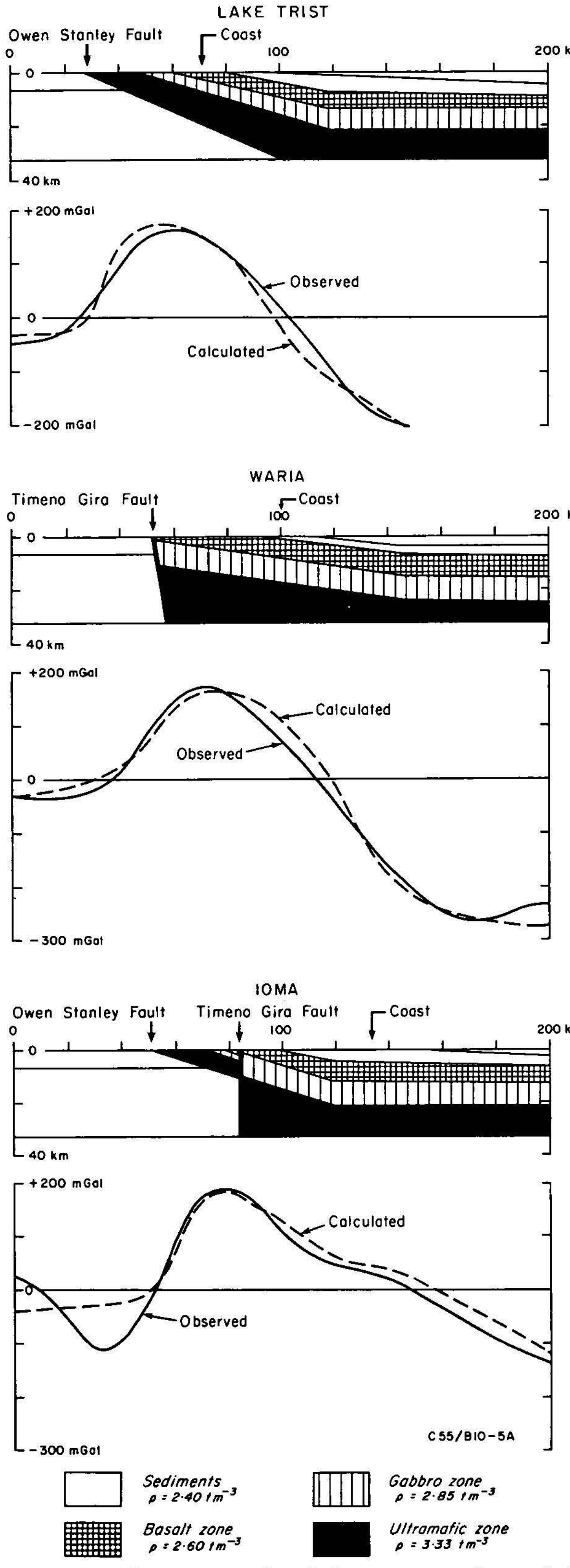


Figure 4. Gravity models of three cross-sections of the Belt. The observed gravity profile Bouguer on land and free air at sea is shown by the heavy line.

The magnetic anomaly on all sections was initially assumed to be caused by the remanent magnetisation of the basaltic layer, which was itself assumed to continue seaward from the Belt under a negligible sediment cover. However, a thick sedimentary layer offshore from the Belt was indicated by the gravity models and the magnetic layer had thus to be made deeper in this area. A good fit could be obtained for the magnetic anomalies over the seaward edge of the belt using the deeper magnetic layer, but even with the automatic fitting technique the match was poor over the area of ultramafic and gabbroic outcrop on the Ioma section.

Two separate modifications (Fig. 6) of the model were tried to overcome this problem, and both improved the fit. The ultramafic and gabbroic components of the Belt were included with a joint susceptibility of 0.12 SI units, and the angle of inclination of the remanent vector was changed to −60°. The −60° inclination model was used to test the idea that the Belt was formed well south of its present location. The higher angle of inclination leads to an improved fit between the observed and calculated profiles for the Waria section. In addition the adoption of this angle of inclination reduces the discrepancy between models of the basalt layer derived from the two different potential fields (i.e., gravity and magnetic) although the agreement is still not good for the Ioma section.

Timing of emplacement and shearing

The timing of emplacement and shearing cannot be determined directly: it is inferred from geological evidence and from plate-tectonic reconstructions. Potassium-argon ages indicate that the gabbroic layer is late Jurassic (150-147 m.y.) and that the basaltic layer is mid-Cretaceous (116 m.y.) (Davies & Smith, 1970). Maastrichtian (69-66 m.y.) planktonic foraminifera have been found in marls associated with the basaltic layer. These data suggest that the crust now exposed in the Papuan Ultramafic Belt was formed during Late Jurassic to mid-Cretaceous times and was still at oceanic depths at the end of the Cretaceous. However, the Maastrichtian fossils were found in the extreme southeast of the Belt; if thrusting was east-west it is quite

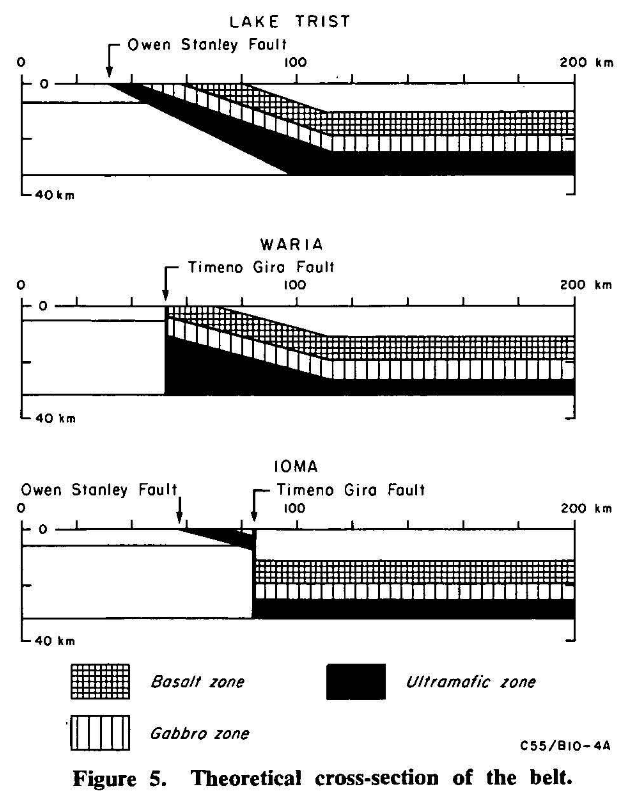


Figure 5. Theoretical cross-section of the belt.

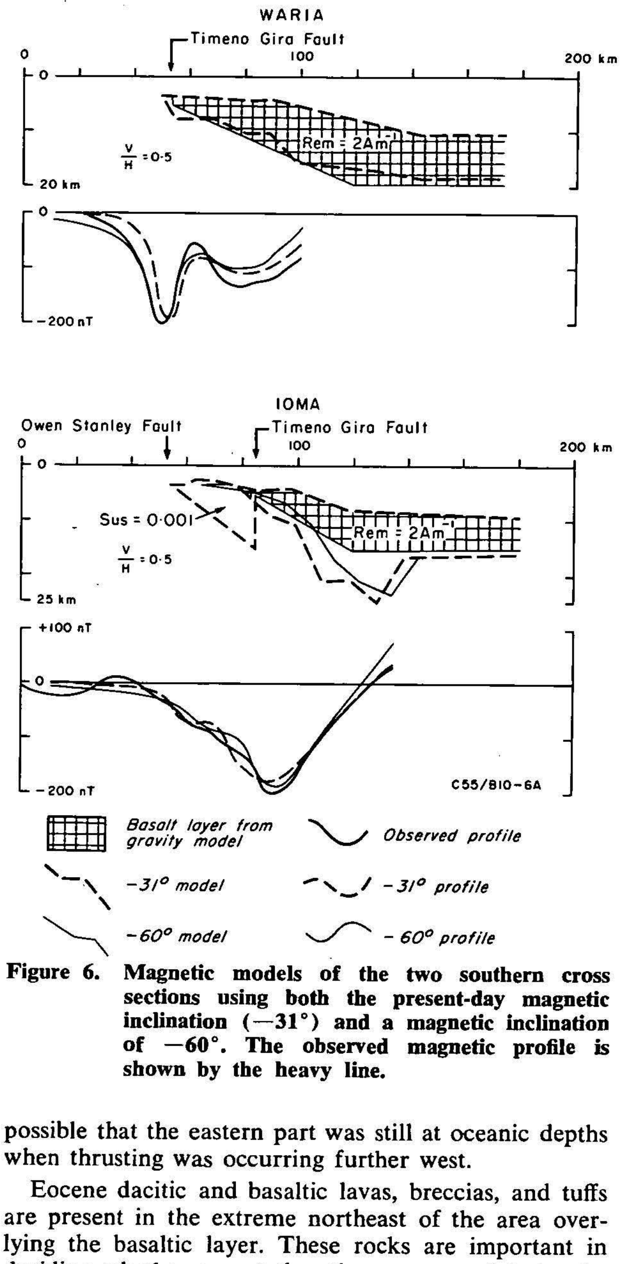


Figure 6. Magnetic models of the two southern cross sections using both the present-day magnetic inclination (—31°) and a magnetic inclination of —60°. The observed magnetic profile is shown by the heavy line.

possible that the eastern part was still at oceanic depths when thrusting was occurring further west.

Eocene dacitic and basaltic lavas, breccias, and tuffs are present in the extreme northeast of the area overlying the basaltic layer. These rocks are important in deciding whether or not thrusting was complete by the Eocene. Their general description suggests that they were laid down in shallow water and, if this is so, thrusting was essentially complete by the Eocene. Thrusting, therefore, is most likely to have occurred at some time between the late Crétaceous (65 m.y.) and the Eocene (55 m.y.), although it remains possible that it occurred somewhat earlier.

A potassium-argon age of 50-55 m.y. has been determined for tonalites which intrude the gabbro-basalt contact, and ages of 52 ± 1 m.y. and 42 ± 4 m.y. have been obtained for metamorphic amphiboles near the basal thrust (Davies, 1977). Isotopic dating gives minimum ages only, and these dates could reflect a tectonic event during the period 65-55 m.y. (Paleocene).

Plate-tectonic reconstructions indicate that the Australian Plate started to move north about 55 m.y. (Weissel & Hayes, 1972), and that the Pacific Plate has been moving west since at least the Cretaceous (Chase, 1971). The relative motion between Australian and Pacific Plates across the Melanesian area can be determined from their poles of rotation with respect to the Antarctic Plate, which is assumed fixed. The present and stage poles (past poles) given by Chase (1971) for the Pacific Plate, and Weissel & Hayes (1972) for the Australian Plate, have been used to compute both the present and past relative motions (Table 2). The location of the plate boundary has changed with time as the plates moved; this change (Table 2) was calculated from the Australian-Antarctic poles of opening given by Weissel & Hayes (1972). The location for the midpoint, in time, of each stage was used. The exact location of the boundary between the two plates is unknown, but as Melanesia is distant from both poles of rotation, relative motions of the two plates calculated for any point in the region will give a representative figure for the whole region.

From Table 2 it can be seen that during most of the Tertiary the relative vector across the Melanesian region has been in a northeast-southwest direction. This vector is approximately at right angles to the direction of compression required to produce the thrusting and shearing of the Belt; consequently these events are unlikely to have occurred later than 55 m.y. Plate-tectonic reconstruction thus support the idea that the Belt was overthrust and sheared before 55 m.y., at a time when the westward motion of the Pacific Plate was the main tectonic force in the area. Overthrusting and shearing were most probably contemporaneous events.

Before the Australian Plate started to move north, its northern margin—along which the thrusting took place—was some 30° of latitude south of its present position. The inclination of the magnetic vector appropriate to this latitude is about —60° and the improved fit of the —60° inclination magnetic models therefore provides further evidence that thrusting occurred prior to 55 m.y.

Lateral extent of the thick Cretaceous crust

The northeastern boundary used for the reconstruction of the belt (Fig. 1b) was chosen for convenience, and does not represent the true offshore extent of thick

Stage	Duration m.y. B.P.	Epoch	Location of mid stage points	Australian Antarctic vector Mag (cm/yr)	Australian Antarctic vector Azimuth	Pacific Antarctic vector Mag (cm/yr)	Pacific Antarctic vector Azimuth	Australian-Pacific vector Mag (cm/yr)	Australian-Pacific vector Azimuth
present			3°S 142°E	7.2	11.6°E of N	11.2	8.6°N of W	13.6	23.2°S of W
I	0-10	Present-L. Miocene	6.2°S 141.4°E	7.2	10.7°E of N	11.0	8.6°N of W	13.3	24.0°S of W
II	10-21.2	L. Miocene-E. Miocene	12.3°S 138.6°E	5.0	17.6°E of N	4.8	10.6°N of W	7.4	32.0°S of W
III	21.2-29.2	E. Miocene-M. Oligocene	16.0°S 136.2°E	4.7	5.3°E of N	5.6	3.9°N of W	7.4	36.0°S of W
IV	29.2-38.2	M. Oligocene-L. Eocene	19.5°S 138.3°E	5.9	15.3°W of N	7.5	3.1°S of W	8.5	46.2°S of W
V	38.2-55	L. Eocene-Paleocene	27.0°S 143.5°E	8.2	23.1°W of N	6.8	1.9°S of W	8.5	65.5°S of W

Table 2. Direction and magnitude of rotation vectors for the Australian and pacific Plates for five epochs during the Tertiary. Vectors have been calculated for a point representative of the position of the Melanesian region during each epoch.

Cretaceous oceanic crust. Gravity models along the Waria traverse show that a simple seaward extension of the Cretaceous crust without any change in thickness adequately explains the presence of the extensive free-air anomaly low (Huon gravity low) offshore from the Belt. Gravity modelling by Finlayson & others (1977) along much the same line as the Waria traverse, but extended to New Britain, shows that there is no change in crustal thickness across this part of the Solomon Sea, and indicates that the Cretaceous oceanic crust extends to the New Britain Trench.

The thick Cretaceous crust would therefore seem to form the floor of the western Solomon Sea and to extend north to the New Britain Trench. The eastern Solomon Sea is, however, of normal oceanic thickness (Furomoto & others, 1966) and it is assumed that a gradational boundary exists between the two parts (Fig. 1a). This boundary would be expected to coincide approximately with the edge of the Huon Gravity Low (Fig. 2). Gravity modelling along the traverse marked Finlayson & others, 1977 in Fig. 1a, which crosses this boundary, shows a gradual thinning of the crust in this area.

The western limit of the thick Cretaceous oceanic crust in this region is probably defined by the small gravity high (Fig. 2) which strikes north across the eastern end of the Huon Peninsula. This high possibly represents an overthrust similar to those further south, but in this case being buried under a thick sequence of later sediments. A substantial pre-Miocene fault, the Murambe-Kwama Fault (Robinson, 1973, 1976) follows the trend of the gravity high, and the existence of the fault and the gravity high, both striking across present structural and geological trends in the area, supports the idea of a buried north-south-striking overthrust in this region. Small outcrops of ultramafic rock are known from this area (Robinson, 1976).

If the Cretaceous crust extends under the Huon Peninsula, then it is likely that it moved to that position by sinistral movement along an easterly extension of the Ramu-Markham fault zone. The offset between the Murambe-Kwama Fault—which represents the approximate location of a buried north-south overthrust—and the most northerly section of the Owen-Stanley Fault—which is a north-south-striking thrust—is about 55 km. The offset is regarded as occurring along a northwest-southeast-trending left-lateral fault (Fig. 1a) similar to the Gira Fault. On land this fault coincides with the Ramu-Markham fault; offshore it probably extends along the north coast of the Papuan Peninsula, possibly as far east as 151°E (Fig. 1a). A north-south seismic-reflection profile across the northern edge of the Trobriand Platform (Finlayson & others, 1976) shows a series of normal faults in the area of the proposed Ramu-Markham extension. The name Solomon Fault Zone is suggested for this fault.

Major gravity and magnetic anomalies similar to those over the Papuan Ultramafic Belt are associated with the northern edge of the Trobriand Platform, which is probably underlain by the thick Cretaceous crust. The gravity anomaly associated with this feature extends east to Woodlark Island and is thus much more extensive than is shown by Finlayson & others (1977, fig. 11). The northern part of the Trobriand Platform was separated from the Cretaceous crust farther south by rifting, which started in early Miocene times (Tjhin, 1976) but which has been most active during the last 3 m.y. (Luyendyk & others, 1973).

Davies (1977) has suggested that an island arc existed to the northeast of the Papuan Ultramafic Belt during the Eocene, and that this arc now forms the floor of the western Solomon Sea and the northern part of the Trobriand Platform. He likened the gravity anomaly over the northern Trobriand Platform to that over New Britain rather than to that over the Papuan Ultramafic Belt. No rocks older than Miocene are known from this area, so the nature of the northern Trobriand Platform remains speculative. However, a number of factors militate against the western Solomon Sea being a submerged island arc.

Topography over present-day island arcs is generally very rugged; in contrast the topography of the western Solomon Sea is fairly subdued—much of it appears to be fault-controlled (Finlayson & others, 1976, fig. 2). It seems unlikely that a uniform thickening could have been produced by island-arc-type plutonic activity, as suggested by Davies (1977). Finlayson & Cull (1973a, b) have shown that the crustal structure of the New Britain arc is extremely complicated, in sharp contrast to the uniform structure of the western Solomon Sea. An active island arc immediately to the north of the Papuan Ultramafic Belt should produce a fairly thick sequence of volcanogenic sediments and lavas on the belt. There is a small area of Eocene volcanogenic sediments in the extreme northeast of the Belt, but in general Cretaceous to Eocene sediments are lacking in the area.

Isostatic compensation of the Papuan Ultramafic Belt and western Solomon Sea

The models shown in Figure 4 were constructed to elucidate the general structure within the crust. No structures below 32 km were included in the model. However, the area includes both high mountain ranges and deep ocean basins. Evidence from other parts of the world shows that these feaures should be isostatically compensated: the mountains by a downward bulge in the base of the crust, and the ocean basin by an upward bulge. St John (1970) has published an isostatically corrected gravity map of New Guinea, computed on the assumption that the crust was 30 km thick and that all topographic masses were compensated by a low density root below this level. His map shows positive isostatic anomalies over the Owen Stanley Ranges with amplitudes of 10-20 mgals, suggesting that the range is close to isostatic equilibrium but slightly undercompensated (i.e., the crustal mass is larger than expected). However, St John's calculation of isostatic anomalies over the Papuan Ultramafic Belt is invalid, as the large excess of mass in the upper crust completely masks the isostatic anomalies. This mass excess and the topography associated with it would be expected to be close to compensation if the adjacent Owen Stanley Ranges are compensated.

A model along the Lake Trist section was constructed, including compensating masses for the topography and mass excess in the crust. The introduction of compensating masses at the base of the crust depressed the general level of the gravity field. The density of the basaltic and gabbroic layers had therefore to be increased and the dip of the contact between the ultramafic rocks and the Owen Stanley metamorphics had also to be increased. However, the basic structure found from the uncompensated model was not altered. Compensated models for the Waria and Ioma sections were not constructed, but there is less dense material in the upper crust on these sections and less modification should be required to produce compensated models.

The calculated gravity anomaly over the compensated model is less than the observed anomaly. This indicates that the Papuan Ultramafic Belt is slightly undercompensated, and in this respect it is the same as the Owen Stanley Ranges. Some of the discrepancy between the observed and calculated anomalies in this area is probably also due to the use of a density 2.67 tm^{-3} for the Bouguer correction. The rocks in this area probably have a density of 3.00 tm^{-3} or more and the Bouguer correction applied would therefore be too low, producing a Bouguer anomaly which was too high.

Offshore from the Belt the type of compensation should change from a low density root to a high density antiroot. However, it was found impossible to model the very deep gravity low offshore if any sort of high density antiroot were present. This area therefore appears to be overcompensated (i.e., the crustal column is too light). The isostatically corrected anomaly map produced by St John (1970) shows that much of the area northeast of the Ramu-Markham Fault is overcompensated, and so the overcompensation in the western Solomon Sea follows this trend. The area is thought to be an active plate margin. It is therefore quite conceivable that it is being held out of compensation by the interaction of the two plates. This mechanism could also account for the slight undercompensation which is found in the Owen Stanley Ranges and the Papuan Ultramafic Belt.

Acknowledgements

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Evidence of evaporite minerals in the Archaean Black Flag Beds, Kalgoorlie, Western Australia

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Sediments with ankerite pseudomorphs after gypsum, and possibly also after anhydrite, have been recognised in the greater than 2675 ± 35 m.y. old Black Flag Beds at Kalgoorlie, Western Australia. This indicates the presence of hypersaline brines during the deposition or diagenesis of these sediments. Recognition of the carbonate pseudomorphs after sulphates supports the conclusion of other recent workers that sedimentary sulphate deposition occurred during the Archaean; added to the growing list of such occurrences, this contradicts earlier conclusions that there are no significant sedimentary sulphates older than about 1000 m.y.

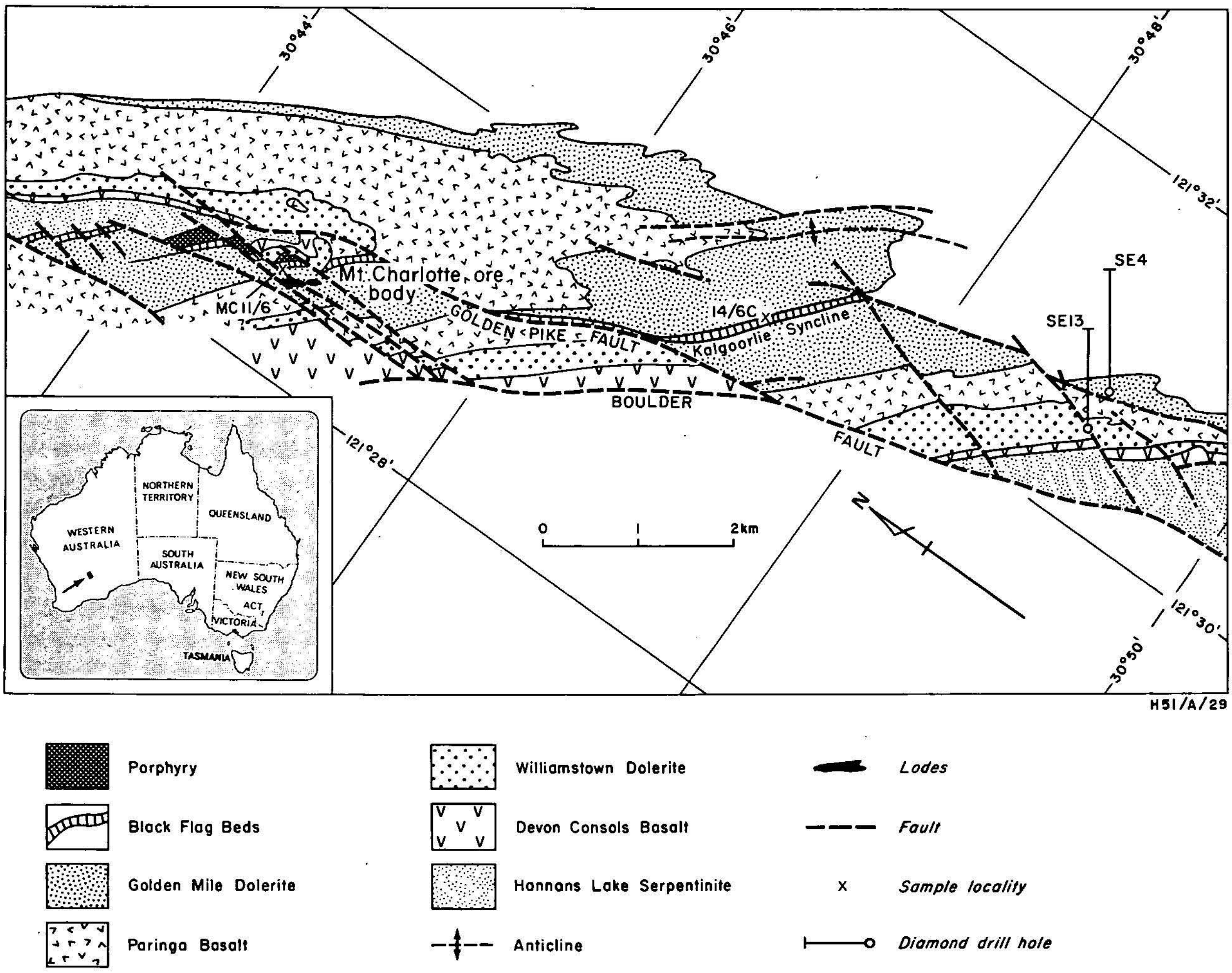

Figure 1. Geological plan of the Kalgoorlie area (after Travis & others, 1971).

Ankerite pseudomorphs after evaporite minerals have been found in sediments from the Archaean Black Flag Beds at Kalgoorlie, Western Australia. The find is significant, as it places restrictions on interpretations of the depositional environment of the previously little studied Black Flag Beds, and supports the conclusion of other recent authors that sedimentary sulphate deposition occurred during the Archaean.

The Black Flag Beds are the uppermost unit of the Kalgoorlie Succession (Table 1), which has been dated by Turek & Compston (1971) as greater than 2675 ± 35 m.y. old. Previous workers (Woodall, 1965; Travis & others, 1971; Travis & Woodall, 1975) have described the Black Flag Beds as a series of acid to intermediate tuffs, agglomerates, flows and intrusions, feldspathic sediments and minor argillites. Detailed study of the Black Flag Beds from both the mining area and an area southeast of the Golden Mile (Golding, 1978) revealed numerous facies changes within the basal part of this unit—on both a local and a regional scale—with tuffs and volcanic breccias pre-

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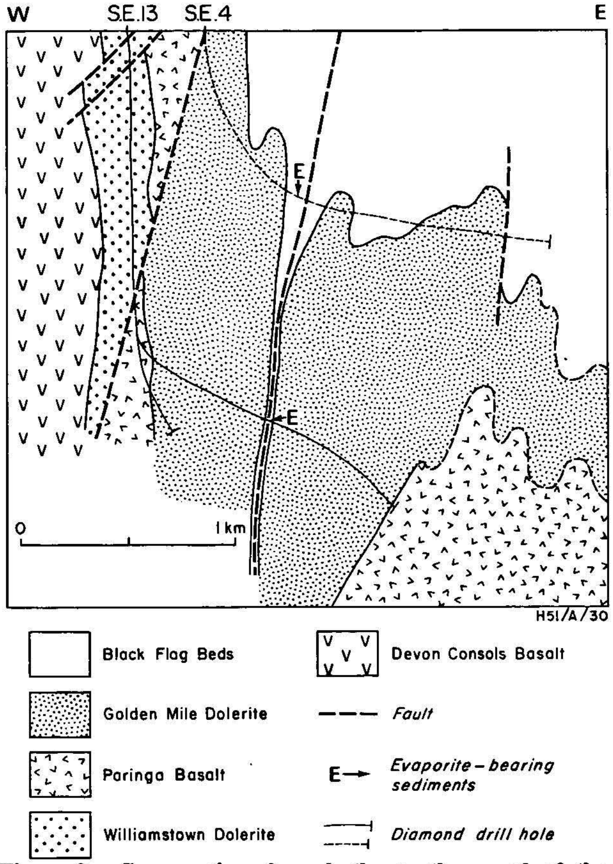


Figure 2. Cross-section through the southern end of the Kalgoorlie goldfield (after Travis & others, 1971). Positions of evaporite-bearing sediments are marked.

Rock Unit	*Approximate thickness (m)*	*Dominant lithology (pre-metamorphism)*
Black Flag Beds	3000(?)	Acid to intermediate tuffs, agglomerates, flows and intrusions; feldspathic sediments; minor argillites
Golden Mile Dolerite	up to 800	Differentiated mafic sill
Paringa Basalt	300-900	Tholeiitic basalts in upper portion, some magnesian basalts in lower portion; minor interflow shales and tuffs
Williamstown Dolerite	up to 300	Layered mafic to ultramafic sill
Kapai Slate	1-5	Graphitic shale, chert and felsic tuff
Devon Consols Basalt	60-100	Magnesian basalts, usually pillowed
Hannans Lake Serpentinite	300-500	Ultramafic flows, chiefly serpentinized peridotites

Table 1. Kalgoorlie stratigraphic succession (from Travis & Woodall, 1975).

dominating in the east and southeast, and fine-grained sediments predominating in the central part of the field. The Kalgoorlie succession has been isoclinally folded and subjected to low-grade regional metamorphism of the quartz-albite-epidote-biotite subfacies of the greenschist facies (Travis & others, 1971; Glikson, 1971). The metamorphic grade decreases to the east of Kalgoorlie (Binns & others, 1976). Figures 1 and 2 illustrate the structure of the Kalgoorlie area, as interpreted by Travis & others (1971). Sediments and pyroclastics of the Black Flag Beds have been tightly infolded and faulted into the Kalgoorlie Syncline (Fig. 2); all pseudomorph-bearing sediments found to date at Kalgoorlie have come from within this structure.

Massive pseudomorph-bearing sediments were intersected by DDH SE 4 drilled by Western Mining Corporation to the southeast of the Golden Mile (Figs. 1 and 2). These sediments are associated with black shales, laminated shales, carbonate-bearing sediments and intraclast breccias.

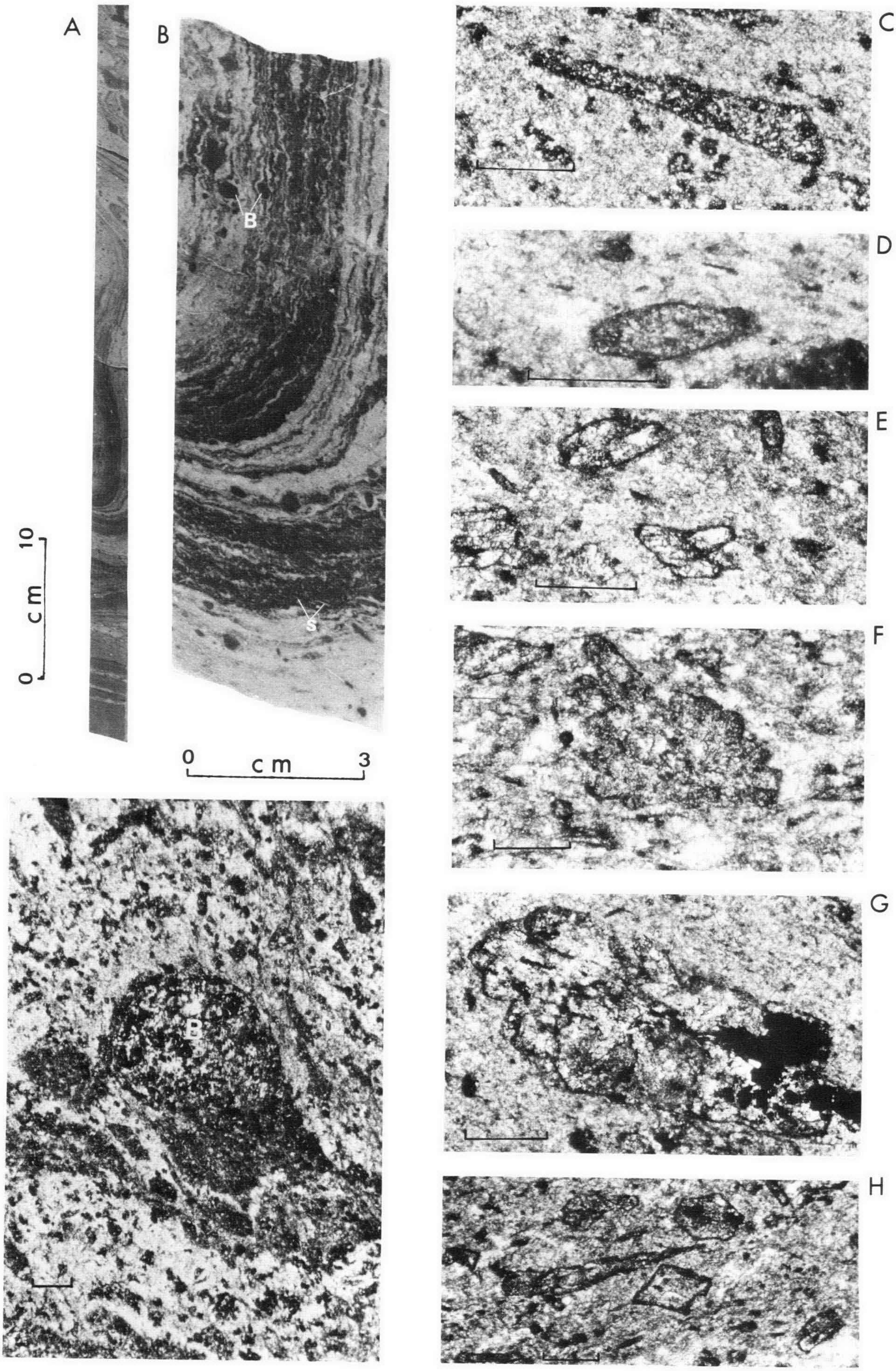
Most of the pseudomorph-bearing sediments are massive quartz-sericite-albite-carbonate rocks, but in places the sediments are laminated. An interval of well laminated sediments with disrupted bedding is illustrated in Figure 3, A and B. The disruption of the bedding is considered to be the result of minor soft-sediment slumping, as the bedding on each side of this section is parallel to that in the overlying and underlying shales.

Both the massive and laminated rocks contain numerous ankerite bodies, which are interpreted as pseudomorphs after sulphates. Some elongate, tapered carbonate forms (Fig. 3C) are of uncertain origin although they resemble anhydrite crystals illustrated by Gill (1977). Forms considered typical of gypsum include lenticular bodies (Fig. 3D) similar to gypsum described by Masson (1955) from the Laguna Madre mudflats of southwest Texas (and to artificially precipitated gypsum; Cody, 1976), interpenetrating twins (Fig. 3E), rosettes (Figs. 3F and G) and crystals with pseudo-hexagonal cross-sections (Fig. 3H).

Most of the Kalgoorlie pseudomorphs are too poorly preserved to allow the measurement of interfacial angles, but one example with a symmetrical hexagonal cross-section has angles of 55.2° and 68.0°, closely comparable with those of gypsum (55.87° and 68.30°); these measurements were made as illustrated by Walker & others (1977). Such angles are unlike those of aragonite, which has a regular hexagonal cross-section. The only other group of minerals with similar angles to gypsum is the amphibole group in which $110 \wedge 1\bar{1}0$ is approximately 56°. However, whilst the elongate forms could conceivably resemble pseudomorphs after amphibole, the twinned grains and rosettes are unlike amphiboles and are typical of gypsum. Furthermore, the low metamorphic grade and low Fe and Mg contents of the rocks (Table 2) are incompatible with the presence of amphiboles of metamorphic derivation; although primary hornblende is a constituent of some

Figure 3. Sediments with sulphate pseudomorphs from the Black Flag Beds. Scale bar on all photomicrographs is 0.5 mm long.

A. Disturbed laminated sediments. B. Small section of A. Light coloured sulphate pseudomorphs (s) are just visible in the dark coloured laminations. Note disruption of the laminae by albite-carbonate nodules (B). C. Elongate tapered carbonate body of uncertain origin. D. Lenticular carbonate pseudomorph after gypsum. E. Carbonate pseudomorphs after gypsum, including interpenetrating twins (lower centre). F. Carbonate pseudomorphs after a rosette of gypsum crystals. G. Carbonate (light) and sulphides (black) after a rosette of gypsum crystals. H. Elongate and blocky pseudomorphs after sulphates. Note pseudohexagonal outline of one of the pseudomorphs. I. Albite-carbonate nodule (B) in laminated sediment.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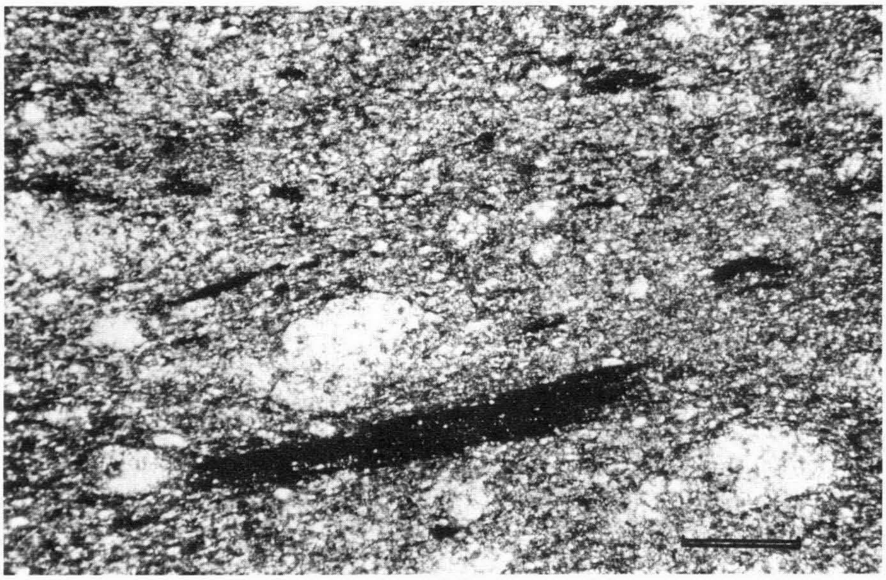

Figure 4. Possible pseudomorphs after small anhydrite nodules in a mud-flake conglomerate. Scale bar is 2 mm long.

of the volcanic rock fragments in pyroclastics from the Black Flag Beds, no hornblende has been observed as a constituent of any of the crystal tuffs or sediments. The only other volcanic products which could resemble the elongate pseudomorphs are glassy volcanic fragments or achneliths (Walker & Croasdale, 1972) and flattened pumice fragments. However, these products do not resemble the lenticular, twinned and rosette forms which are typical of gypsum. It is concluded that the pseudomorphs are after sulphates.

The pseudomorphs now consist dominantly of ankerite, but quartz and sericite are also present in some. A few of the pseudomorphs consist partially of sulphides (sphalerite, chalcopyrite, and gersdorffite) (Fig. 3G) but in general the sulphur content of these rocks is very low (Table 2), indicating significant loss of sulphur. Similar carbonate pseudomorphs after sulphates are described from Proterozoic rocks of the McArthur Basin by Walker & others (1977).

Numerous dark, rounded blebs consisting of carbonate and albite interrupt the bedding in the laminated sediments (Figs. 3B and I). Albite grains and aggregates up to 7 mm long are also scattered through the massive rock. The origin of these is uncertain.

The sulphate pseudomorphs and carbonate-albite nodules are set in a fine-grained matrix consisting of quartz, sericite, carbonate and a little albite. In the well-laminated rocks, a small amount of carbon occurs within the darker laminations. Chemically, there is little difference between the various rock types containing pseudomorphs (Table 2). The sediments probably contain a significant volcanogenic component, but some of the silica could have been introduced later.

Similar sediments were found in the core of the Kalgoorlie Syncline in DDH SE 13 (Fig. 2). In this intersection, the sediments are very silicified, but the pseudomorphs after sulphates can still be recognised. Silicified sediments with elongate pseudomorphs apparently after sulphates were also found in the mining area on the No. 6 level of the Chaffers mine (14/6C), and as fragments in an intraformational conglomerate from the No. 11 level of the Mount Charlotte mine (MC11/6). Irregular carbonate patches in a mud-flake conglomerate from the No. 6 level of the Chaffers mine may be after small anhydrite nodules

	1	*2*	*3*
SiO_2	64.22	69.33	61.31
TiO_2	0.36	0.27	0.32
Al_2O_3	14.96	11.75	13.43
FeO	2.09	2.07	3.16
MnO	0.04	0.03	0.04
MgO	1.77	1.31	1.89
CaO	3.64	3.51	5.41
K_2O	2.92	2.30	2.25
P_2O_5	0.19	0.15	0.18
Na_2O	2.86	2.55	2.83
Ba	0.08	0.05	0.06
S	0.02	0.01	0.05
CO_2	6.1	4.9	7.5
H_2O	1.0	1.2	1.1
Total	100.03	99.31	99.53

Table 2. Analyses of evaporite-bearing sediments from Kalgoorlie.

1. 2832/4 Massive grey sediment with pseudomorphs after sulphates. 2. 2849/4a Light coloured sediment with pseudomorphs after sulphates. 3. 2849/4b Well laminated sediment with pseudomorphs after sulphates.

(Fig. 4); anhydrite nodules of similar size have been described by Gill (1977) from the Salina A-1 sabkha cycles of the Michigan Basin.

The regional S and C isotope study of the Yilgarn carried out by Donnelly & others (1977) did not find any evidence of sedimentary carbonate deposition or sulphate reduction. Unfortunately, no evaporite-bearing sediments were included in that study and only cross-cutting carbonates were analysed from the Black Flag Bed sediments. The anhydrite from the mineralised Golden Mile Dolerite sample analysed by Donnelly & others is of hydrothermal origin, taking the form of a discordant veinlet; this is quite distinct from the sulphate pseudomorphs in the Black Flag Beds, which have definite crystalline shapes.

The association of evaporite-bearing sediments at Kalgoorlie with intraclast breccias, pyritic bituminous shales, carbonate-bearing sediments, and shales with lenticular bedding has been interpreted by Golding (1978) as suggestive of a shallow water to subaerial environment similar to that of modern tidal flats. Certainly, the former occurrence of evaporites must be explained in any palaeo-environmental model for these sediments.

At Kalgoorlie, conditions favourable for sulphate formation may have existed for a long period of time, as the sediments are thick (with an intersection of 41 m in DDH SE 4 giving a true thickness of about 38.5 m); if the evaporite-bearing sediments in DDH SE 13 are equivalent to those in DDH SE 4, the deposit has a width of at least 1000 m. Since possible evaporite-bearing sediments have been found in the mining area as well as in drilling to the southeast of the Golden Mile, the deposit may extend for many kilometres.

When Cloud (1968, 1976) formulated his model for the evolution of the atmosphere and hydrosphere, the oldest well authenticated extensive sulphate evaporites were in the late Precambrian Bitter Springs Formation of central Australia. Cloud took that to indicate that before Bitter Springs time the atmosphere was insufficiently oxygenated to allow the precipitation of sulphate minerals from the oceans. This interpretation has been objected to on theoretical grounds, because little or no free oxygen in the atmosphere is required for the precipitation of sulphates (Lambert 1978; Lambert & others, 1978). It is now apparent that extensive sulphate deposition occurred prior to Bitter Springs time, in the Proterozoic (e.g. Walker & others, 1977), and in the Archaean (Viljoen & Viljoen, 1969; Perry and others, 1971; Hickman, 1973; Heinrichs & Reimer, 1977; Radhakrishna & Vasudev, 1977; Vinogradov & others, 1977; Dunlop, 1978; Dunlop & Groves, 1978; Lambert, 1978; Lambert & others, 1978). The occurrence of ankerite pseudomorphs after gypsum in the Black Flag Beds reported here supports previous interpretations which have indicated that sedimentary sulphates were deposited in Archaean as well as later times.

Acknowledgements

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Zircons in the granitic rocks of southeastern South Australia

J. B. Colwell

A morphological study of zircons from the granitic rocks of southeastern South Australia has been made using reduced major axes, length/breadth ratios, colour, zoning and other features. In general, zircons in the biotite adamellites (Encounter Bay and Kingston areas) are more elongate, less euhedral and freer of inclusions and zoning than those occurring in the other granitic rocks of the region. Results suggest that the biotite adamellites are genetically related (they share the same zircon characteristics) and that they are unrelated to the more widespread 'eastern zone' of hornblende-biotite granite, quartz porphyry and hornblende and biotite microgranite. The rocks in the eastern zone share very similar zircons and appear to be related.

Introduction

Granitic rocks, which are apparently of Early Palaeozoic age, occur as discontinuous outcrops over a wide area of southeastern South Australia, and extend into Western Victoria (Fig. 1). They form part of the Padthaway Ridge, an area of shallow basement separating sediments of the Otway (Gambier Embayment) and Murray Basins. The rocks, which consist of rhyolite, quartz keratophyre, hornblende/biotite granite and microgranite, and biotite adamellite, have been described by Mawson and co-workers (1943 to 1945), Henstridge (1970), and Rochow (1971). They form a western zone extending along the eastern flank of the Mount Lofty metamorphic belt (including the "Encounter Bay Granites") and an eastern zone, the "Murray Bridge Granites", extending southeastwards from the metamorphic belt to Dergholm in western Victoria (Milnes & others, 1977).

This paper presents the results of a morphological study of zircon from the granitic rocks, work undertaken as part of an overall investigation of heavy minerals occurring in the late Cainozoic sediments, and igneous and other source rocks of the region (Colwell, 1979). The purpose of this paper was to see whether the various granitic rocks of the region could be correlated on the basis of their zircon fraction, the term correlated implying derivation from the same or closely related magmas.

A number of workers (for example, Poldervaart, 1956; Larsen & Poldervaart, 1957; Spotts, 1962) have noted that the relatively early formation and short range of crystallisation of zircon in granitic rocks commonly results in crystals of the mineral, in a single intrusive body or related intrusive bodies, having similar morphological or elongation characteristics. These characteristics can be relatively easily measured in grain mounts and the results expressed statistically as length/breadth ratios and reduced major axes.

Method of study

Samples were collected from outcrop at the sites shown in Figure 2. Samples were crushed in a jaw crusher and disc grinder, and the material split and sieved into –88, 88 to 177, and 177 to 500 micronsize fractions. Heavy minerals were separated from each fraction in bromoform. In all cases, most of the zircons were found to lie in the –88 micron fraction. Grain mounts of this fraction were prepared using "De Pex" as the mounting medium. Conclusions based on this work make the assumption that the –88 micron zircons are representative.

Two-hundred unbroken zircon crystals were examined for each sample, using a polarising microscope fitted with a mechanical stage. Colour, habit, zoning, overgrowths and inclusions were noted. Length and breadth measurements were made using a micrometer ocular, and the following values calculated: mean length ($\bar{x}$) of the zircons, mean breadth ($\bar{y}$), standard deviation of the length (s_x), standard deviation of the breadth

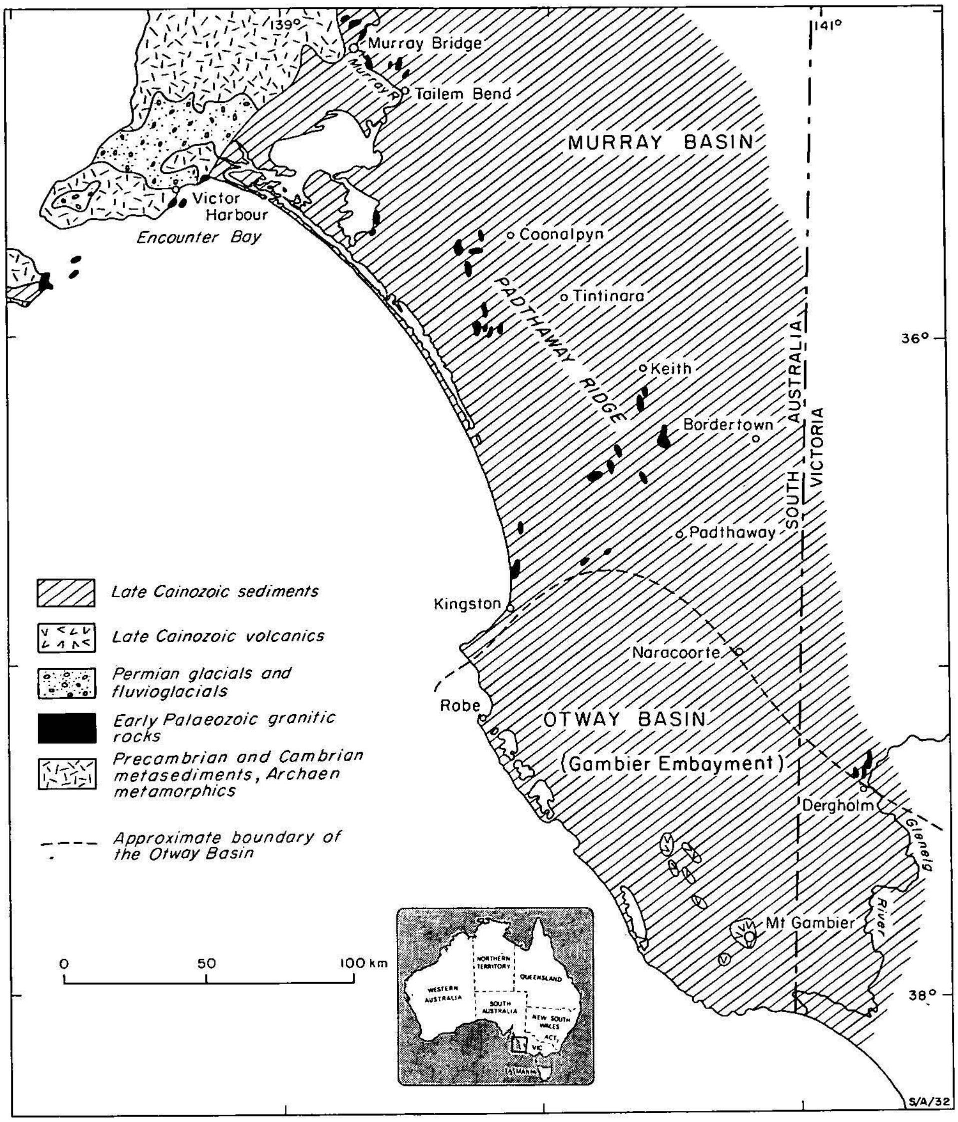

Figure 1. General geology of southeastern South Australia.

(s_y), and slope (a) (equal to s_y/s_x). Mean length and mean breadth were plotted as a point on a length-versus-breadth graph, and a straight line with a slope equal to the ratio (a) drawn through the point ($\bar{x}$, $\bar{y}$). This line is the reduced major axis (Kermack & Haldane, 1950) and represents the relative growth trend or pattern of crystal growth for the zircon sample (Spotts, 1962).

Comparisons of the plots of the reduced major axes were made by visual inspection. Further tests of elongation characteristics were made using elongation-frequency diagrams.

In all, 14 samples were used in the study: 4 of biotite adamellite, 2 of quartz porphyry, 3 of hornblende-biotite granite, 3 of biotite granite and 2 of hornblende granite. Petrographic descriptions of the samples are available (Colwell, 1976). A sample of rhyolite (Papineau Rocks) failed to provide sufficient zircons for study.

Results and discussion

Typical zircons extracted from the rocks are shown in Figure 3. Reduced major axes are plotted in Figure 4, and frequency polygons of zircon elongation in Figure 5. Colour, zoning, and other features of the zircons are summarised in Table 1.

The zircon data separate the biotite adamellite from the other rocks. In general, zircons in the biotite adamellite are more elongate, less euhedral, and freer of inclusions and zoning than those occurring in the other rocks. Crystal growth trends (indicated by reduced major axes) differ in a similar fashion. No separation is apparent between the biotite adamellite of the Encounter Bay area and occurrences farther south, northeast of Kingston.

The quartz porphyry (Mount Monster) and the hornblende, hornblende-biotite, and biotite granite and microgranite have the same or very similar zircon characteristics. This is consistent with general chemical and mineralogical similarities noted for the rocks by Mawson and coworkers (1943 to 1945).

Overall, the zircon study supports the view expressed by Mawson and coworkers, and others that the biotite adamellite bodies throughout the region are genetically related, and that they are unrelated to the more widespread 'eastern zone' of genetically related hornblende-biotite granite, quartz porphyry, and hornblende and biotite microgranite.

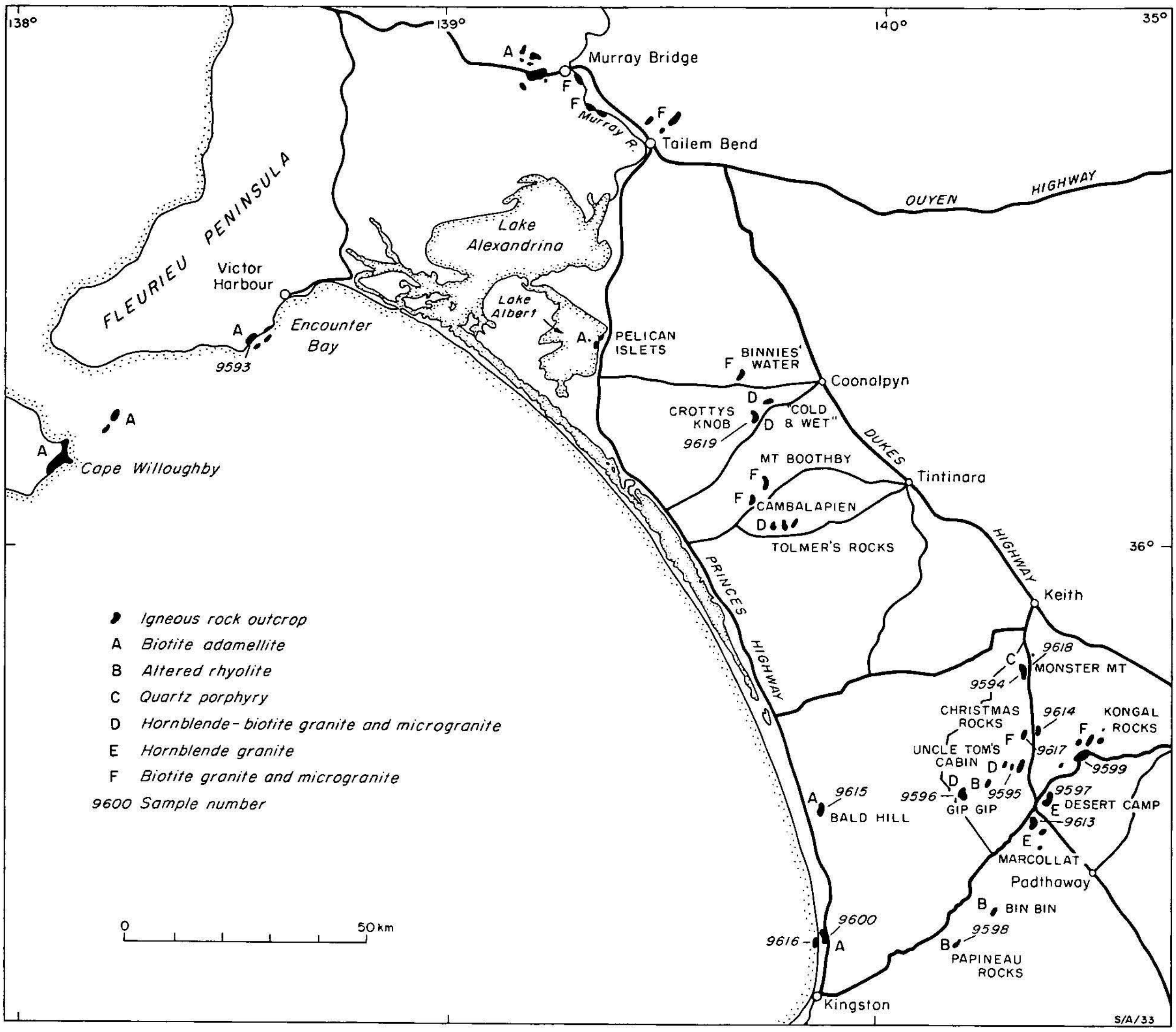

Figure 2. The location of the samples and the distribution of the different types of igneous rock.

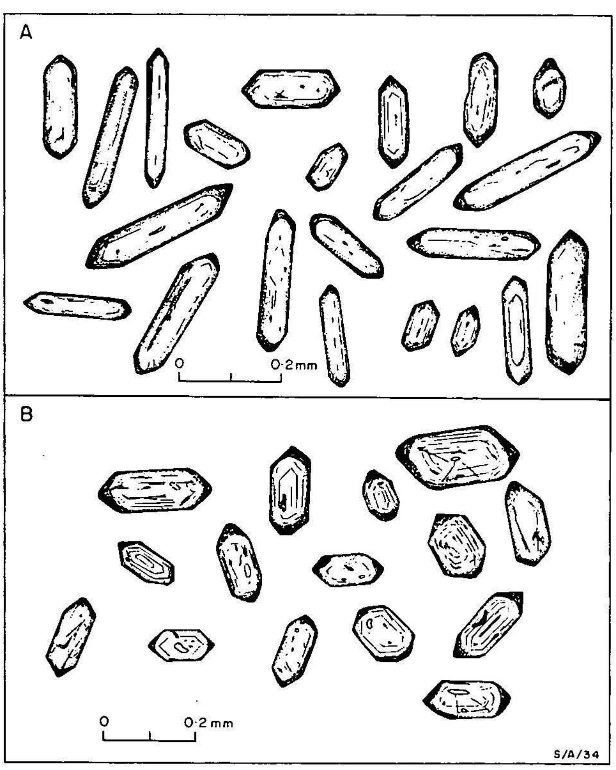

Figure 3. Typical zircons in (A) the biotite adamellite, and (B) the hornblende/biotite granite, microgranite and quartz porphyry.

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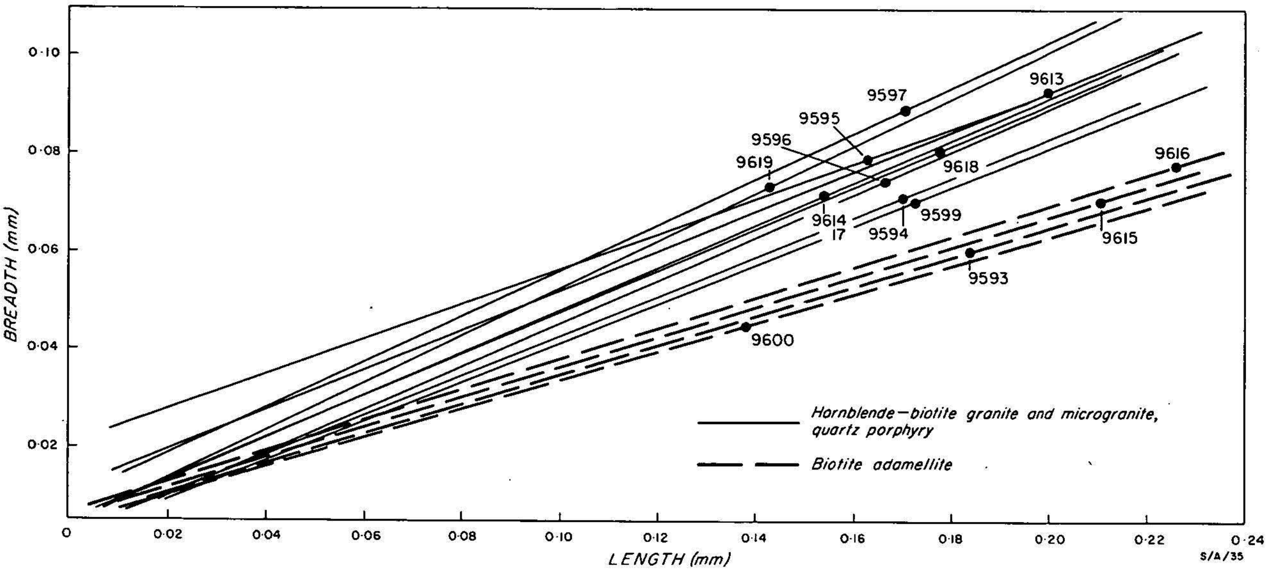

Figure 4. Reduced major axes.

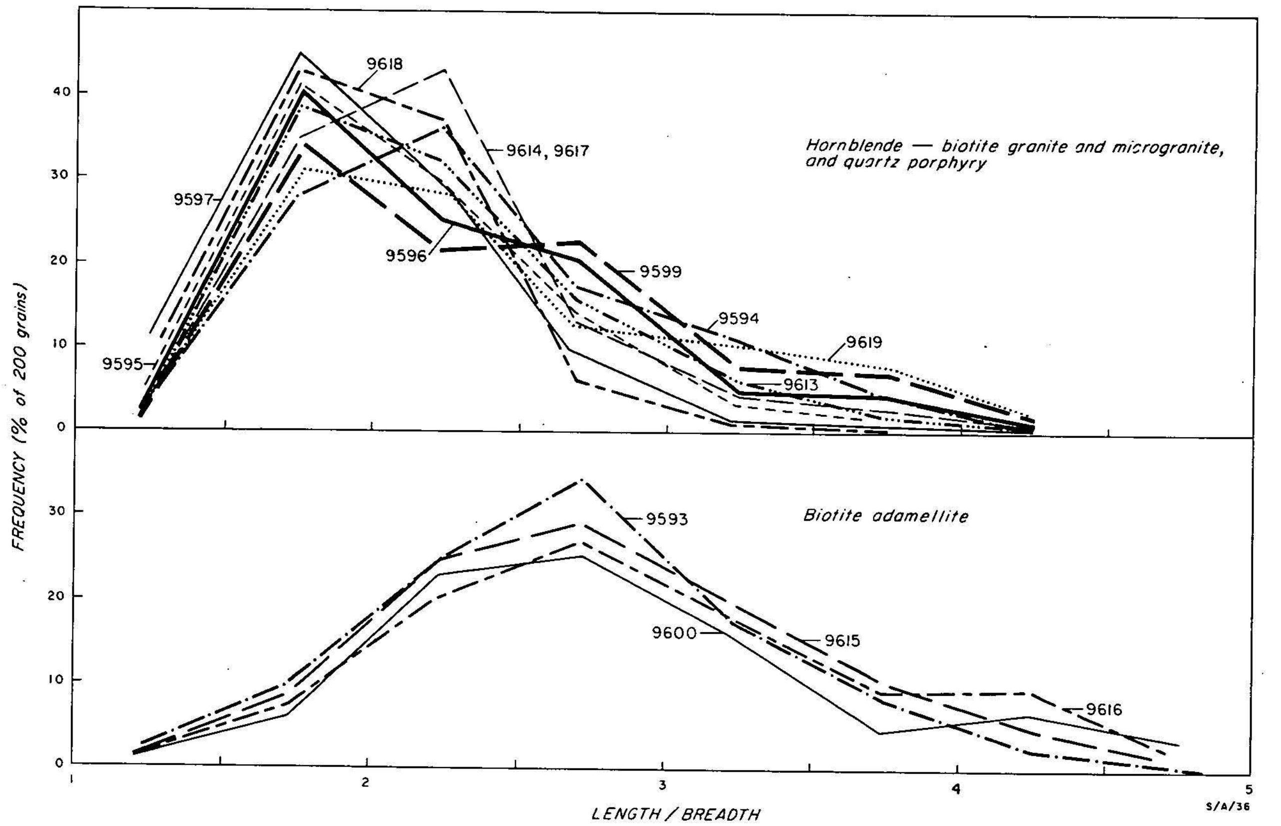

Figure 5. Frequency polygons of zircon elongation.

Sample No.		9593	9616	9600	9615	9594	9618	9595	9619	9596	9599	9614 & 9617	9597	9613
Rock type		Adamellite	Adamellite	Adamellite	Adamellite	Quartz porphyry	Quartz porphyry	Hornblende-biotite granite	Hornblende-biotite granite	Hornblende-biotite granite	Biotite granite	Biotite granite	Hornblende granite	Hornblende granite
		%	%	%	%	%	%	%	%	%	%	%	%	%
Colour	Colourless	90	94	97	95	96	98	93	94	92	92	93	97	99
	Pale yellow	10	6	3	5	4	2	5	6	5	8	7	3	1
	Pink	0	0	0	0	0	0	2	0	3	0	0	0	0
Habit	Euhedral	61	54	57	58	88	90	86	77	81	80	84	78	78
	Slightly rounded	31	32	30	27	10	9	12	16	15	14	14	18	18
	Corroded	8	14	13	15	2	1	2	7	4	6	2	4	4
Zoning	Moderate to well-developed	36	35	37	32	61	54	64	58	27	44	69	67	22
	Poorly developed	10	23	19	24	20	10	12	14	30	12	16	15	23
	Absent	54	42	44	44	19	36	24	28	43	44	15	18	55
Zircon overgrowths	Present	8	3	2	1	2	1	3	2	5	4	3	2	1
Inclusions	Absent or very rare	44	31	37	45	13	10	9	21	16	22	15	29	31

Table 1. Features of the zircons. %: percentage of the 200 zircons examined in each sample. Sample numbers are prefixed 7563.

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BMR Journal of Australian Geology & Geophysics, 4, 1979, 77-78

The late Cainozoic evolution of the Carpentaria Plains, North Queensland: a discussion

R. J. Wasson[1]

Grimes & Doutch (1978) have identified five predominantly alluvial stages in the depositional landforms of the Carpentaria Plains. In the light of the factors which they consider to be the most likely controls on the evolution of the plains, they suggest a correlation between these stages, sea-level, climatic changes, and other depositional sequences in Queensland.

The five-stage sequence

The Carpentaria Plains have been divided by Grimes & Doutch into: 1, the large alluvial fans to the north of Normanton; 2, the southern alluvial plains of less distinctive plan form. The five stages have been identified by essentially morphological criteria, although some lithological criteria have been used.

Fan surfaces have been mapped previously using both lithological and morphological criteria (Hunt & Mabey, 1966; Hooke, 1967; Denny, 1967; Wasson, 1974), but the apparent ease with which these surfaces have been mapped does not necessarily mean that a direct equation can be made between morphological and lithostratigraphic units. The five stages may be well represented morphologically in the Carpentaria fans, but, as Grimes & Doutch admit (p. 104), substages may exist within each of the 'units'. Without detailed stratigraphic data it is difficult to determine how realistic the five-stage sequence is.

A most significant aspect of the paper is the correlation of the fan sequence with that in the southern plains. Grimes & Doutch note (p. 108) that the divisions between the putative five stages in the southern plains are not as clear as they are in the fans. Sediments of Stages 1 and 2 in the fans are both ferruginised and silicified, and Stage 3 is only ferruginised. However, although it is evident that Stage 1 in the southern plains is ferruginised, it is not at all clear from the text of the paper that Stage 2 is ferruginised. In the discussion of the Wondoola Plain it is suggested that it formed mainly during Stage 3, but there is no mention of ferruginisation in its case.

The doubts expressed by Grimes & Doutch, and the absence of crucial criteria in the southern plains, suggests that no faith can be placed in the correlation of sequences between the two areas. Moreover, it is possible that substages do exist within the major units which may or may not correlate with each other. The work of Williams (1970, 1973), Wasson (1977), and Schumm (1968) demonstrates some of the problems which arise in the stratigraphy of alluvial fans, both small and large.

Interpretation

The factors which have controlled the development of the alluvial plains are seen by Grimes and Doutch as: tectonic; sea-level fluctuations; climatic fluctuations. In a brief discussion the authors appear to remove earth movements from consideration, except in a few cases, and look for correlations between the five stages and other events in Queensland which are likely to be related to fluctuations of either sea-level or climate.

According to the authors, the Claraville Cycle of deposition, which produced the Carpentaria Plains, began by upwarping of the margins of the Karumba Basin. Movement has occurred sporadically in the uplands ever since, and Grimes & Doutch acknowledge some role for tectonism in the evolution of the plains when they note: 'Depositional stages during the Claraville Cycle may have been in part a response to this mild tectonism, . . .' (p. 103). However, they largely disregard tectonism when attempting correlations. The authors also note that: '. . . the fans themselves display no obvious signs of dislocation . . .' (p. 103). This argument is faulty, for earth movements within the catchments of the fans, which are noted by the authors, would undoubtedly affect the behaviour of the fans. Furthermore, Hooke (1972) has provided a most detailed account of the segmentation of fans as the result of tilting, with no signs of dislocation of the fans. Tectonic activity cannot be ruled out as a factor worthy of further consideration.

Sea-level is considered an important factor because of its role as ultimate base level for the streams of the area. The correlations in the authors' figure 6 are partly related to sea level, and climatic fluctuations are viewed as significant, but there is little discussion of these factors.

Any interpretation of alluvial fan histories must follow from an understanding of the mode of evolution of fans. Alluvial fans invariably accumulate in spatially discontinuous units; that is, deposition occurs on one part of a fan's surface for a period of time, but does not occur over the entire surface at any one time (Bull, 1977). As sediments accumulate in one area the difference in elevation between that area and the surrounding surface increases. Eventually the area of deposition shifts to a lower part of the fan surface either by a shift in the fan-building streams or by a change in the paths of debris flows (cf. Hooke, 1967). Gole & Chitale (1966) have documented the gradual change in the location of streams on the Kosi Fan, a fan in northern India of comparable size to the largest in the Carpentaria Plains, with a gradient of 1:1200 as against 1:1800 (average) on the Carpentaria fans. Schumm (1977) has used the Kosi Fan as an example of accumulation by lateral shifting of the locus of deposition.

Lateral shifting on fans is often accompanied by periodic longitudinal changes in the locus of deposition. Fan-head trenching shifts the locus down-fan (Hooke, 1967; Wasson, 1974), and back-filling of the trench moves the locus up-fan. The factors which are thought to control fan-head trenching are numerous, and there are two schools of thought. Firstly, there are those who argue that fan dissection and construction of fan units (in the terminology of Grimes & Doutch) is the result of a fundamental change in regime. This change can be caused by either long-term climatic change (Lustig,

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1965) or diastrophism, particularly tilting and faulting (Bull, 1964; Hooke, 1967).

Secondly, there are those who argue that fan-head trenching is part of the general fan accumulation regime. Within this regime Bull (1964), and Beaty (1974) argue that extreme floods are responsible for entrenchment (in the sense of Wasson, 1977) and a shift in the locus of deposition. Hooke (1967) presents experimental evidence that entrenchment occurs as the result of alternations between debris flows and water flows, a mechanism which is unlikely to be relevant to the Carpentaria Plains. Eckis (1928) suggests that continued downcutting within a fan's catchment will inevitably result in the dissection of the fan's head and therefore a shift in the locus of deposition. This mechanism must operate on a much longer time scale than either of the others discussed, and may be relevant to the Carpentaria Fans. Finally, Weaver & Schumm (1974), and Schumm (1977), have argued from experimental results that fanhead trenching can occur as the gradient of the fan head increases during deposition to a threshold value beyond which entrenchment will occur (cf. Schumm, 1973).

It is clear from this summary that the progress of fan accumulation is complex and that the isolation of controlling factors is difficult. At a more general level, the evolution of alluvial landscapes has presented more problems of interpretation than many other landscapes (cf. Haynes, 1968; and Flint, 1971). The relationship between climatic fluctuations and changes in the alluvial landscape is at the moment very difficult to determine.

There is no reason to believe that the Carpentaria Fans are less complex than other fans. The stratigraphic approach is the only one which will provide information suitable for isolation of the controlling factors. The discussion of stratigraphy and controls by Grimes & Doutch is inadequate as a contribution to the study of alluvial fans.

Quaternary correlations

Without absolute time control of the Carpentaria Plains sequence, either radiometrically or palaeomagnetically, the only basis for correlation with other sequences in Queensland is proof of a relationship between identified stratigraphic units and a regional control such as climatic or sea-level changes. The five-stage sequence is in some doubt, and so we cannot be sure that the evolution of the plains has been in response to any regional control. This is sufficient reason not to pursue regional correlations.

Furthermore, it has not been demonstrated that the evolution of the fan units has been in response to climatic or sea-level changes, rather than the tectonic factor or to variables inherent in fan building.

The conceptual basis for the correlations of figure 6 is not established and the correlations must be suspect. Grimes & Doutch conclude that '. . . the depositional sequence in the Carpentaria Plains could provide a base for correlating similar sequences throughout much of Queensland' (p. 111). This is unlikely and, in the light of the complexity of alluvial landscapes generally, and alluvial fans in particular, it is more likely that the Carpentaria Plains may provide one of the worst bases for correlation with other sequences.

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BMR Journal of Australian Geology & Geophysics, 4, 1979, 79-81

The late Cainozoic evolution of the Carpentaria Plains, North Queensland: reply

K. G. Grimes[1] *& H. F. Doutch*[2]

Wasson (1979) discusses our interpretation and correlation of the fluvial sequence in the Carpentaria Plains (Grimes & Doutch, 1978). His main criticisms are:

(1) he considers that we have insufficient 'stratigraphic' data to substantiate our division of the fans sequence into five stages and to support our correlation of these stages with the sequence in the southern plains;

(2) he criticises the brevity of our discussion of climatic and sea levels controls on the evolution of the fans and our 'apparent' exclusion of tectonic effects, and would seem to consider that the shifts in the depositional areas of the fans are not related to environmental changes;

(3) in view of his doubts concerning the above points, he considers that the regional correlations which we suggested are suspect and that the fans provide a poor basis for attempting such correlations.

We will discuss these points in sequence.

Recognition of the five-stage sequence

Wasson considers that our five stages in the fans area may not be realistic litho-stratigraphic units, as their recognition lacks 'detailed stratigraphic data'. On similar grounds he rejects our sequence in the southern plains and the correlation between the two areas.

First, we would like to point out that the main criterion in the erection of any stratigraphic unit is mappability. The fan stages are eminently mappable and therefore demand an explanation. We have mapped them and attributed their development to regional environmental controls.

The traditional concepts of lithostratigraphy have evolved mainly from the study of older lithified rocks. Younger, undisturbed deposits, on which the original depositional surfaces and landforms are still recognisable, allow a morpho-stratigraphic approach to their mapping which is not possible in the study of older rocks (Frye & Williams, 1962). In the Carpentaria Plains the morpho-stratigraphic approach is much more useful than lithostratigraphy, which is hampered by variations due to provenance at the regional scale, and by environmentally controlled facies (and a lack of subsurface data) at the local scale. Note that the Wondoola and Claraville Beds, though distinctive lithostratigraphic units, are contemporaneous and their differentiation has no time-stratigraphic significance. It is time-stratigraphy which has been our main concern in the recognition and attempted correlation of the five stages in the Carpentaria Plains.

The morpho-stratigraphic (and time-stratigraphic) criteria used in delineating the fans units are summarised in Grimes & Doutch (1978, p. 103-5). The two most important criteria used in assessing time relationships were: firstly, the cross-cutting boundaries of the younger units over the older units (analogous to unconformities in a lithostratigraphic approach) and, secondly, variations in the degree of preservation of the fluvial landforms, which allow relative ages to be assessed. When one sees well preserved, and sharply defined, sandy levee and channel deposits in one area, and more degraded, ill-defined features of the same type in an adjoining area it seems reasonable to assume a significant period of time has passed between the deposition in the separate areas.

The stratigraphic relationships between the younger fan stages and the late Pleistocene and Holocene beach ridges also assist both in recognising the different fan stages as distinctive events and in providing some absolute control on their ages (p. 104-5).

Wasson is distracted by our reference to the possibility of substages in the two older fan units. We have no evidence for the existence of such substages but merely suggested the possibility because of the apparently long time span covered by these two stages. Even if subsidiary stages did exist they would no more detract from the significance of the major stages than would the existence of 'members' detract from the recognition of a 'formation' in a lithostratigraphic approach.

In all we feel that these criteria are sufficient to justify the division of the fans into five morpho-stratigraphic units, and to allow the recognition of their time-stratigraphic relationships.

The southern plains

We indicated that the divisions in the southern plains were not as clear cut as in the north. This is partly because of the over-riding effect of provenance controlled lithofacies: it is difficult to compare morphological features on a sand plain with those on a mud plain. None the less stratigraphic relationships can be observed, using both lithological and morphological criteria. The basis of our subdivision and its correlation with the fans sequence is discussed in our original paper (p. 108). We will enlarge here on some specific points raised by Wasson.

The stage 2 deposits of the Claraville Plain are ferruginised. The deposits of the Wondoola Plain are not, but this reflects lithological control rather than an age difference. The heavy black soils of the Wondoola Plain would not lend themselves to ferruginisation as readily as the sandy soils of the Claraville Plain, and the fans.

One relationship not mentioned in our text (though it is shown on our figure 3) is that the Claraville Plain abuts the Gilbert Fan. In this area, on the basis of the degree of degradation of old levee ridges and channel sands, the surface of the Claraville Plain appears similar to the G_2 fan, but older than the G_3 fan where the sand ridges of the old levee banks are better preserved.

1. Geological Survey of Queensland, 2 Edward St, Brisbane 4000; published with the permission of the Under Secretary, Queensland Department of Mines.

While a relative sequence can be recognised in the southern plains we must admit that apart from correlation of the Claraville Plain with stage 2 and the Millungera Plain with stage 3 (p. 108), our comparison between the two areas is based mainly on the relative sequence and the assumption that the sequences in the two areas are both the result of regional environmental controls (i.e. tectonism, climate, and sea level). This assumption is discussed below.

Factors controlling the fans sequence

Wasson proceeds with a discussion of the factors controlling the evolution of fans in general, and criticises our interpretation of the Gilbert and Mitchell Fans.

He describes two schools of thought concerning the periodic longitudinal changes in the locus of deposition. One relates these shifts to environmental changes, while the other regards them as part of the general fan regime. Wasson does not state which, if either, of these he supports. However, his aim appears to be to demonstrate that our five fan stages could be the result of local, spontaneous effects rather than environmental changes of regional significance. We do not feel that he has demonstrated this to be so.

Some caution is needed in relating studies on alluvial fans to the sequence in the Carpentaria Plains. When we commenced work on this paper we were in some doubt as to the applicability of the term 'fans' to such large, flat-lying features. Previous workers had referred to them merely as plains (Twidale, 1966, and Galloway & others, 1970) or as 'deltas' (Warner, 1968). However, the term had been used for similar features in the Riverine Plain of NSW (Butler & others, 1973) and on the Ganges Plain (Douglas, 1977, p. 171).

The 'fans' of the Carpentaria Plains are extreme end members of the group of features known as alluvial fans. Anstey (1965, cited in Cooke & Warren, 1973, p. 175-6) has listed dimensions of fans ranging in length from less than half a kilometre up to 25 km with a mode between 2 and 5 km; and the gradients from 1:5 to 1:120 (mode between 1:20 and 1:60). By contrast the 'fan' segments in the Carpentaria Plains are up to 170 km long and have much lower gradients (1:1800 on average). Further, their evolution is spread over a longer period of time than many of the classical fans described in the literature.

While there are obvious analogies between the fans of the Carpentaria Plains and 'classical' fans, the difference in both spatial and temporal scales means that one should be careful of assuming too great a similarity of process. Thus Wasson's references to debris flows and the disruption of small fans by extreme floods would have little relevance to the Carpentaria Plains.

A study of the 1974 flooding of the plains (Simpson & Doutch, 1977) showed a situation different from that of classical fans. Large areas of the plains were covered by slowly moving sheets of water, much of which was clear, ponded rainwater. Silty water from the uplands was confined to the major channels and erosion and deposition appeared to be limited in extent.

Wasson refers to the swinging of the streams across the surface of the large Kosi Fan as an example of a continuous process resulting from aggradation of successive channels (Gole & Chitale, 1966). Doubtless this type of relatively rapid movement was responsible for the deposition of the individual fan stages in north Queensland. However, the shifts in the overall locus of the fans, which distinguish our fan units, is a different matter.

He also refers to continual regular downcutting of a fan's catchment, and its head. This, if regular, would produce a gradual shift in the locus of a large fan. But a gradual shift is unlikely to produce a small number of discrete segments such as occur in the Carpentaria Plains, where it would seem that downcutting has been in several stages separated by periods of stability or aggradation.

While in a small relatively steep fan a downstream shift in the locus of deposition may sometimes be attributable to a short-term random fluctuation in discharge or some other factor, a shift of the magnitude found in the large fans of the Carpentaria Plains is more likely to represent a long-term variation in the environment. In fact, most authors have attributed fan-head entrenchment and downstream shifts of the depositional area to environmental changes—either tectonic or climatic.

The main problem lies not so much in the attribution of the fan stages to environmental controls as in identifying the specific controls involved.

Tectonism, climate, or sea level?

The three controls would seem to be tectonic effects, climatic effects, and, for these fans, sea-level changes. We discussed the influence of tectonism on the first two stages, but not on the last three stages. The history of tectonism is deduced indirectly from the episodes of dated volcanism in the uplands from which the source rivers flow. The bulk of the volcanics were formed in the Pliocene and early Pleistocene, and for the earlier fan stages tectonism is probably a major environmental cause of deposition. On page 110 we indicated tentative correlations between stages 1 and 2 and dated periods of volcanicity. By the late Quaternary, when the last three stages formed the volcanism (and hence presumably the tectonism) was less widespread.

On the other hand we have evidence for late Quaternary sea-level changes and there are climatic models related to these changes. As tectonism appears to have become less active we have assumed that climate and sea level were the more important controls for the last three stages. As the climatic models of Webster & Streten (1972), and Nix & Kalma (1972), are closely linked to sea-level changes these two effects cannot easily be separated. The most direct influence of sea level would have been on stages 3 and 5, which are related to high stands.

We do not intend to add to the current discussions on the relationships between specific climates and periods of stream incision and aggradation. Our paper aims to describe the morpho-stratigraphic sequence in the Carpentaria Plains and to suggest that the sequence is related to environmental changes of regional significance (tectonism, climate, and sea level). Tectonism and climate would have affected all the streams rising in the Einasleigh and Burdekin Uplands, which forms a major drainage divide. Sea level would not have affected the inland streams but would have contributed to some extent in the depositional history in the other streams. Thus regional correlations are to be expected.

Wasson is correct in stating that more detailed studies are needed in order to make a detailed assessment of the geological and geomorphological development of the plains, and of the specific environmental changes involved. Our paper, and Wasson's discussion of it, should provide a starting point for such studies. However, we do not feel that his criticism detracts from our general conclusions: that a five-fold sequence exists and that similar sequences may be expected elsewhere in the region. To reiterate our final statement in full, 'it would seem that, *with care*, the depositional sequence in the Carpentaria Plains could provide a base for correlating similar sequences throughout much of Queensland'. The likelihood of these correlations decreases with distance from the starting point, but further field studies in these areas will provide the most appropriate test of the hypothesis. It should not be abandoned because of misgivings about the significance of alluvial fan development elsewhere in the world.

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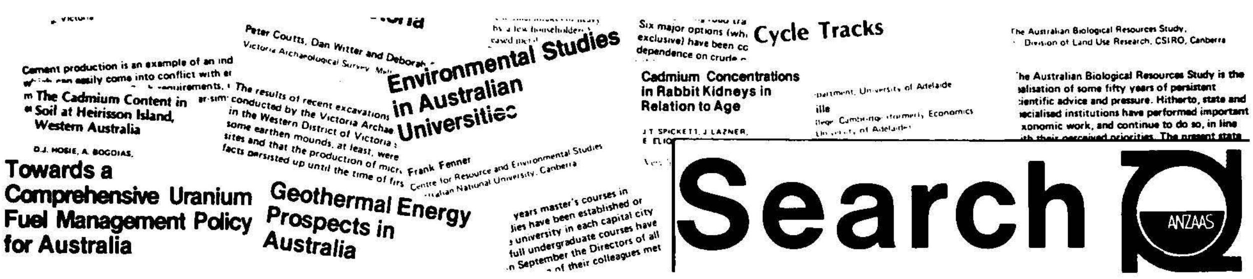
Corrigenda

We would like to take advantage of this reply to correct a few errors in the text and figures of the original paper:

p. 102; 'C, Coleman Fan;' has been omitted from the reference to figure 1. On the face of the figure the '. . . old land surfaces' stipple pattern has been omitted from an area to the west of the 40 m contour and extending north from the Holroyd R. to the edge of the figure. The unlabelled physiographic region between the Mitchell and Alice Rivers is the Alice Plain and is probably of stage 2 age. The unstippled eastern part of the Croydon Plain (Cr.P.) should carry the 'erosional areas' stipple of the area to the east. p. 104; column 2, line 20, should read 'All surfaces of stage 3 or older, . . .' p. 108; column 2, line 15, the last word should be 'There'. p. 111; column 1, line 21, should read '. . . which are now ferricreted;'.



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