AUSTRALIAN AND NEW ZEALAND GEOMORPHOLOGY GROUP

SECOND CONFERENCE

Broken Hill, 8-12 July 1984

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Sunday 8 July. Afternoon/evening arrive Broken Hill and settle in.

Monday 9 July. Themes: semi-arid and arid landscapes; soil and landscape.

9.15	Introductory welcome (John Chappell)
9.30	N.A. Abraham. Geomorphology and land systems of the far northwestern corner of NSW.
10.00	J.M. Bowler. SLEADS in progress: Rates and groundwater processes in Arid Zone Evolution.
10.30	-- Morning tea --
11.00	J.A. Mabbutt. Sand availability, equivalent sand thickness and logitudinal dune patterns, in the northwestern Simpson Desert, central Australia.
11.30	R.J. Wasson and K. Fitchett. Pattern and process in the continental dunefields of Australia.
12.00	G. McTainsh. Desert loess deposition processes in West Africa.
12.30	-- Lunch --
1.30	R.W. Young. Arid climate or rock control: problems in the interpretation of sandstone terraine.
2.00	A.P. Spate, D.S. Gillieson and J.N. Jennings. Anomalous red quartz sands in certain Nullarbor Plains caves and dolines.
2.30	D.J. Eldridge. Soil distribution on aeolian landforms in southwestern NSW.
3.00	P.J. Walker. Landscape and soil development in the Cobar-Tilpa district, NSW.
3.30	-- Afternoon tea --
4.00	C. Chartres. Soil development in the Barrier Range, western NSW.
4.30	D. van Dijk. Relationships between surface gravels and stages in soil geomorphic planar developments, Tara Plains, southeast Queensland.
5.00	Hope, J. Pleist-Palaeoenvironments on the Darling River.
8.00	CONFERENCE SPEECH. Professor J.A. Mabbutt: Review of present and future directions in arid zone geomorphology in Australia.

Tuesday 10 July. Themes: Denudation, and fluvial geomorphology.

9.00 R. Vertessy. The production of storm flow runoff from a burnt forested hillslope.

9.30 E. Stock and R. Neller. Biophysical responses to hydrology of Blackwater Creek, Central Queensland.

10.00 D.L. Dunkerley. Influence of lithology on solutional processes, Chillagoe karst, north Queensland.

10.30 -- Morning tea --

11.00 J. Field. The characterization of a catchment from stream loads.

11.30 J. Chappell. Holocene sedimentation and river dynamics in the South Alligator and Daly Rivers, N.T. 55 B denudation (each in estuary) channel denudation 2-3 B

12.00 J. Soons. Uplift and erosion on the West Coast of the South Island, New Zealand.

12.30 -- Lunch --

1.30 D. Outhet. Sediment deposition in Lake Wyangala, NSW.

2.00 J. Chappell. Uplift and erosion on the Huon Peninsula, New Guinea.

2.30 W.N. Jenks. Relating form and process in tidal channels: an evaluation of the work per unit area approach from the Manning River Delta, NSW.

3.00 D. Hean. Problems with the application of bedload formulae to NSW coastal rivers.

3.30 -- Afternoon tea --

4.00 G. Nanson. Bedload transport yields determined from meander migration rates.

4.30 R. Warner. Channel changes: adjustments? to what?

5.00 W. Erskine and M. Melville. Recent channel changes and their relevance to river management: a case study of Daisy Arm, Hunter Valley, NSW.

7.00 CONFERENCE DINNER St Peter Hall, Lane St (Chloride + Oxide St)

Wednesday 11 July

8.00 Day excursion amidst the geomorphology of the Broken Hill and Barrier Range region.

8.00 pm Business meeting of our entire Group, to discuss future conferences, group activities, publication and any other business.

Thursday 12 July. Themes: Quaternary, process, and regional and large-scale Geomorphology.

9.00	C. Woodroffe. Hydrodynamics and particulate matter flux in a mangrove basin, Auckland NZ.
9.30	R.W. Galloway. Fossil Ice Wedges in Patagonia and their Palaeoclimatic Significance.
10.00	R.P. Bourman. The Late Paleozoic glaciation of southern South Australia.
10.30	-- Morning tea --
11.00	D. Adamson. Gullying at Belarabon & Mungo in N.S.W.
11.30	M.A.J. Williams. Late Quaternary landscape of west-central NSW as an analogue to the British Permo-Triassic.
12.00	G. Bowman and N. Harvey. Holocene evolution of the LeFevre Peninsula, Adelaide.
12.30	-- Lunch --
1.30	M. Connolly. Deep leads and landscape interpretation in New England NSW, with Victorian comparisons.
2.00	L. Worrall. Sub-basaltic topography, Barrington Tops, NSW.
2.30	C.F. Pain. Landsat imagery and landform mapping.
3.00	P. Wellman. The topography of Australia, its origin, and geophysical control.
3.30	-- Afternoon tea --
4.00	P. Williams. Origin of tower karst in southern China.
4.30	R.W. Blair. Development of natural sandstone arches in southeast Utah.
5.00	G. Speight. The logical framework of historical geomorphology.

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GEOMORPHOLOGY AND LAND SYSTEMS OF THE FAR NORTH-WESTERN CORNER OF N.S.W. - AN OVERVIEW

N.A. Abraham

Soil Conservation Service of N.S.W.
Nyngan

ABSTRACT

The land systems of the far north-western corner of New South Wales have been used as a consistent basis for the systematic description of the geomorphology and for an understanding of the geomorphic history of the area. This paper outlines the physiographic regions and the main controls on landform development.

INTRODUCTION

The land system sheets of Milparinka, Cobham Lake, White Cliffs and Urisino have been published by the Soil Conservation Service of N.S.W. at a scale of 1:250 000 based on the national mapping grid of that scale (Milthorpe (1978), Lawrie and Stanley (1980 a, b,) Milthorpe et al. (1979)). The land systems were defined and mapped following the principles outlined by Christian (1958).

The area has been divided into 10 physiographic regions that have been used for the systematic description of the geomorphology and land forming processes and controls. Geological structure has controlled the broad distribution of the regions, its influence most apparent in the Uplands, where erosional processes are dominant. Tertiary silcrete has had a primary control in a number of regions. The Alluvial Plains are dominated by active depositional processes, with aeolian processes most active in the Sand Plain regions. The outline of the physiographic regions is shown in figure 1.

PHYSIOGRAPHIC REGIONS

Uplands

a) The Barrier Range. This region, in the south-west, consists of the Coco Range and the northern end of the Barrier Range, being formed by uplift of the Euriowie Block. Geological structure is the most evident control on the landforms, producing the characteristic strike ridges and rugged hills.

b) Northern Ranges, Tablelands and Plains. This region comprises the Mt. Browne and Warratta Ranges, Mt. Arrowsmith and the extensive stony tableland and plains of Tertiary silcrete in the north-west. The uplift of the Tibooburra Dome has controlled the broad outline of the region, while the Tertiary silcrete is the primary control on the form of the remnant tablelands and mesas, and contributes the stony mantle of the plains.

c) Central Ranges and Slopes. The most spectacular relief and highest elevation are included in this region. Koonenberry Mountain and the Coturaundee Range are fault-bounded blocks while the undulating slopes and lowlands reflect folding of the Palaeozic basement. The Kara Hills have been identified as a remnant of a previous surface.

d) Stony Tablelands and Plains. These extensive stony plateaux, plains and lowlands in the south-east owe their form mainly to the Tertiary silcrete surface and its subsequent deformation and dissection. This has produced the characteristic landforms of tabular uplands with silcrete caps, steep escarpments and a silcrete mantled pediment linking the uplands with the gibber covered undulating lowlands.

e) Mount Pleasant. Occurring in the far south-east of the survey area, this region comprises rounded sandstone hills and ranges, isolate mesa remnants and sand covered footslopes and sandplains. Geological structure has been the main control on landform development.

Alluvial Plains

f) Bancannia Plains. This region is the topographic expression of the Bancannia Trough, tending north north west-south south east between the Barrier Range and the Central Ranges and Slopes. Active alluvial processes are clearly evident in the outwash plains and flats tending towards terminal floodouts and playas along the axis of the trough. In the eastern section there are extensive sandplains with low sand rises, with drainage from the Central Ranges confined to linear depressions.

g) Bulloo and Cobham Plains. This is the largest region in the survey area and is the topographic expression of part of the Bulloo embayment of the Great Artesian Basin. The landforms of the Cobham Plains consist of stable parallel dunes with alluvial tracts draining the Northern and Central Ranges, terminating in playas and lakes. These plains have been formed by both aeolian and alluvial processes. The Bulloo Plains reflect the combined influence of active alluvial process and relict depositional forms. The Bulloo Overflow is but a remnant of a much larger system.

h) Paroo Plains. This region in the far east of the survey area comprises the channels, extensive floodplain and basins of the Paroo River, with areas of adjacent sandplain. Active process and relict forms are evident in the landforms of the region.

Sand Plains

i) Western Plains. This is part of the Lake Frome embayment of the Great Artesian Basin. The main influence on landform development has been active aeolian processes, forming a sand plain with longitudinal dunes generally aligned east-northeast with confined alluvial tracts, playas and floodouts receiving drainage from the Northern Ranges and Tablelands and the Barrier Range.

j) North-Eastern Sand Plain. This region in the north-east of the survey area comprises slightly undulating sand plains with broad sandy rises. Active aeolian processes have been the primary control on landform development.

HISTORY OF LANDFORM DEVELOPMENT

Under the system of morphogenic regions, Mabbutt (1972) has defined the survey area as arid. This recognises the relationship between the main geomorphic processes and forms of the area and the climatic parameters under which they operate. However, interpretation of the landforms of this area can not be based solely on the prevailing climate and a single resultant morphogenesis but also upon major geologic control and the persistence of forms moulded under earlier contrasting climates.

In the study area, consideration of the following influences is necessary to understand the development of the landforms. Table 1 identifies the controls primarily responsible for the characteristic landforms of each land system.

1) Geological Structure. On the broad scale, geological structure has controlled the outline and form of the physiographic regions. The ranges have been formed by uplift and folding of rocks. The plains are generally the topographic expressions of structural lows in the basement. Depositional landforms are most evident in the plains.

The influence of structural control is most evident in the landforms of the range country, such as the broad cuestas on moderately dipping beds of Nundooka land system, the alignment of strike ridges on steeply dipping sediments in Faraway and Teamsters land systems and structural plateaux on flat lying sediments of Kara Hills land system.

2) Tertiary Silcrete. Much of the uplands, slopes and stony plains are formed on the gently folded silcrete surface referred to by Wopfner (1978) as the "Cordillo Surface". He places the main period of silicification as the mid-Tertiary. The duricrust surface has subsequently been subject to tectonic folding and faulting, forming the structural pattern evident today.

The influence of the silcrete can be most clearly seen in the silcrete-capped tablelands of Quarry View and Questa Park land systems and the mesas of Flat Top land system. The fragmentation of the silcrete has generally been the most significant source of the stone mantle of the slopes and plains, such as Olive Downs and Nuntherungie land systems.

3) Relict Depositional Forms. The unconsolidated Tertiary sediments in the depositional basins formed by the deformation of the Cordillo surface have subsequently been subjected to alternating alluvial and aeolian processes during the Pleistocene period. The lunette is one landform that clearly shows this inherited form, with one mode of formation being the deflation of material from the floor of periodically dry lakes.

Other relief features can be seen in the ancient channels of Tongo land system and the scalded alluvial plains of Conservation, Sturt and Yancannia land systems, whose texture-contrast soils are more in accord with prior stream deposition than active floodplains.

4) Active Depositional Processes. Alluvial processes are most evident in the drainage lines, floodplains and lakes of the Alluvial Plains and in the drainage corridors and terminals of the Sand Plains. Drainage lines such as Yapunyah, Gum Vale and Fowlers land systems only carry flows following heavy localised rains while Paroo land system receives flows from flooding in southern Queensland.

The aeolian processes controlling the development of the Sand Plains are defined as active in this review. However, Wasson (1983) puts the last main period of dune formation at 25 000 to 13 000 BP, during the Last Glacial, in windier conditions than present. A number of carbonate zones have been identified within the longitudinal dunes, suggesting earlier periods of dune formation. Wasson concluded that the longitudinal dunes of the east Strezlecki, which adjoins the Western Plains region were mainly derived from alluvial and lacustrine sediments and the break-down of bedrock.

The longitudinal dunes, particularly of Corner and Gumpopla land systems, show a distinct trend denoting a dominant formative wind from the west-southwest and offering no evidence of a subsequent shift in the pressure-wind system (Mabbutt, 1972). However Wasson (pers. comm) has noted a steepening of the southern flanks indicating a resultant wind more to the north west. There was probably also a reactivation in the period 3 000 to 1 000 BP. Current sand movement is generally restricted to isolated crests.

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FIGURE 1

PHYSIOGRAPHIC REGIONS OF THE FAR N.W. CORNER OF N.S.W.

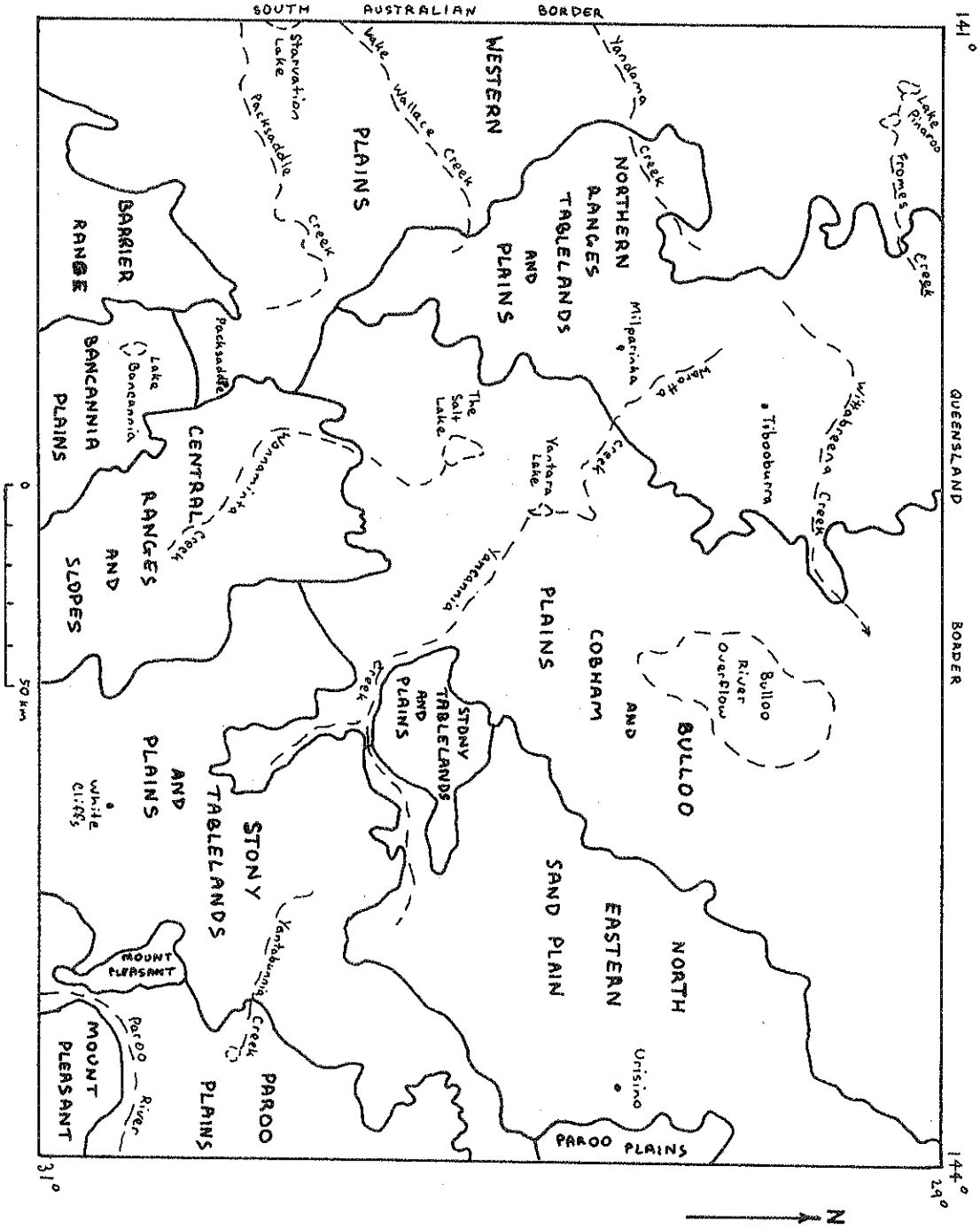

TABLE 1

LAND SYSTEMS AND THE PRIMARY CONTROL ON LANDFORM DEVELOPMENT

Primary Control on Landform Development	Broad Landform Types	Erosional landforms based on: Underlying Geology	Erosional landforms based on: Inherited landforms	Depositional landforms: Alluvial	Depositional landforms: Aeolian
Geological Structure		Faraway Mt. Pleasant Bynguano Nundooka Mt. Arrowsmith Teamsters Wonaminta Mulga Dam Tibooburra Kayrunnera Euramurtie Tekum	Pulchra Cymbric Flood's Creek Onepah Katalpa Nundora Glenhope Oakvale		
Tertiary Silcrete		Mount Pleasant* Nundooka* Kara Hills Ravendale Purnanga Black Stump Tekum* Petita Bancannia	Quarry View Questa Park Pulgamurtie Womparley Flat-top Purnanga* Wonga Lilli Olive Downs Nuntherungie Gap Hills Euramurtie* Petita* Bancannia*	Gum Vale* Fowlers* Yancannia*	
Relict Depositional Forms	Alluvial	Bancannia*		Conservation* Sturt* Tongo* Yancannia* Fort Grey Teurika* Tongowoko	
	Aeolian	Mulga Dam* Tibooburra* Black stump* Euramurtie * Tekum* Petita*	Womparley* Wonga Lilli* Nundora* Glen Hope Oakvale*	Tongo* Paroo* Allandy* Colane* Langawirra Fort Grey* Cobham* Tongewoko*	
Active Depositional Processes	Alluvial			Caloola Yapunyah Narriarra Gum Vale Fowlers Conservation Sturt Tongo Paroo Yancannia Allandy Colane Langawirra Fort Grey* Cobham Teurika Bullagree Paroo Overflow Tongowoko	Rodges* Gumbalara* Nucha* Border* Wonga* Gum Poplah* Corner*
	Aeolian				Rodges Waverly Goodooga Nucha Reola Border Wonga Gum Poplah Klondyke Corner Marrapina Binerah

Note: *indicates that only some of the land units are due to this control.

TABLE 1

LAND SYSTEMS AND THE PRIMARY CONTROL ON LANDFORM DEVELOPMENT

Primary Control on Landform Development	Broad Landform Types	Erosional landforms based on: Underlying Geology	Erosional landforms based on: Inherited landforms	Depositional landforms: Alluvial	Depositional landforms: Aeolian
Geological Structure		Faraway Mt. Pleasant Ryngyano Nundooka Mt. Arrowsmith Teamsters Wonaminta Mulga Dam Tibooburra Kayrunnera Baramurtie Telum	Dulchra Cymbric Floodie Creek Onepah Kalalpa Nundooka Glenhope Oakvale		
Tertiary Silcrete		Mount Pleasant* Nundooka* Kara Hills Ravendale Rumbiaga Black Stump Telum* Petlita Bancannia	Quarry View Quanda Park Baramurtie Wompurley Flat-top Ramonga* Wonga Hill Olive Downs Nundooka* Cap Hills Baramurtie* Petlita* Bancannia*	Gum Vale* Rowlers* Yancannia*	
Relict Depositional Forms	Alluvial	Bancannia*		Conservation* Chard* Tongo* Yancannia* Fort Grey Teurika* Tongowoko	
	Aeolian	Mulga Dam* Tibooburra* Black Stump* Baramurtie * Telum* Petlita*	Wompurley* Wonga Hill* Nundooka* Glen Hope Oakvale*	Tongo* Paroo* Allandy* Colane* Langawirra Fort Grey* Cobham* Tongowoko*	
Active Depositional Processes	Alluvial			Cobola Yapunyah Nundinna Gum Vale Rowlers Conservation Chard Tongo Paroo Yancannia Allandy Colane Langawirra Fort Grey* Cobham Teurika Bullagree Paroo Overflow Tongowoko	Bodges* Gumbalana* Nucha* Border* Wonga* Gum Poplah* Corner*
	Aeolian				Bodges Waverly Goodonga Nucha Beula Border Wonga Gum Poplah Klondyke Corner Marragina Bincarh

Note: *Indicates that only some of the land units are due to this control.

Gullying at Belarabon and Mungo in Western N.S.W.

D. Adamson and M.A.J. Williams

Presently active gullies in stream channels and on hillsides are of no great age. In some cases, a limit on their age is set by living trees of Eucalyptus, Callitris and Casuarina which established within formerly stable stream channels. The living trees also allow the cross-sections of the former stable stream channels to be reconstructed. Living trees, stumps of logged trees, or other features also set age limits for the most recent phase of deposition on alluvial fans below the main gullied zone. The present phase of gullying and outwash probably fits within the time since European settlement.

Earlier phases of erosion of valley fills have occurred, as shown by exhumed channel fills and stone pavements.

DEVELOPMENT OF NATURAL SANDSTONE ARCHES IN SOUTHEASTERN UTAH

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ABSTRACT

The region around Arches National Park and Canyonlands National Park in southeastern Utah displays the highest concentration of natural sandstone arches in the United States and perhaps the world. Rock fins are a prerequisite to arch formation and develop either from closely spaced joints or from canyon wall retreat from opposing sides of adjacent drainages. Factors important in arch formation include 1) the horizontal attitude of sandstone formations, 2) the presence of less competent horizons, 3) the systematic release of stressed rock via exfoliation slabs, 4) the continued weakening of surface materials by hydration and wet-dry cycles, and 5) the erosion of weathered products by wind, running water, and gravity. The arches have probably all evolved within the Quaternary Epoch.

INTRODUCTION

More than 300 natural sandstone arches are found in the semi-arid southeastern Utah, U.S.A., which may represent the highest concentration of this type of landform found anywhere in the world. Most of the arch openings are concentrated within two prominent sandstone formations, the Cedar Mesa Sandstone Member of the Cutler Formation (Permian) and the Entrada Sandstone (Jurassic). This preliminary study is concerned with the factors responsible for the development of these sandstone arches.

PREREQUISITES TO ARCH FORMATION

According to Cleland (1910) a natural arch is not the same as a natural bridge and, therefore, should be treated separately. Bridges span present or former water courses while arches do not. Bridges result from the corrasion and hydraulic action of streams or possibly waves, while arches result from weathering and less intense erosional processes.

The most important prerequisite to the formation of sandstone arches is the development of a thin wall of rock or rock fin. Two types of fin development are recognized, those which evolve from closely-spaced jointing and those which result from headward erosion and valley wall retreat from adjacent drainages.

The best examples of joint-controlled fins are found in the Devil's Garden section of Arches National Park, Utah, where hundreds of joints striking N40°W run parallel to the axis of the Salt Valley Anticline. These are bending moment joints associated with the development and collapse of the Salt Valley Anticline. The anticline has been evolving since Triassic time from the slow upward migration of a ridge of salt (Barrs, 1972). During the Quaternary and possibly Late Tertiary, solution and removal of the salt near the fold axis resulted in the partial collapse of the fold limbs in toward the axis to produce a topographic valley. Because of the brittle nature of the Entrada Sandstone, the limb folding event resulted in the development of open joints narrowing toward the base and rock fins 3 to 6 m wide and a relief of 6 to 40 m (Blair and others, 1976). Some fins run for hundreds of meters without a break.

Many of the joints have separated to produce long narrow passages also several meters across. Sand silt along with spalled rock fragments have filled some of the interjoint passages. In others weathering and stream channel erosion has resulted in the widening of the interjoint areas. These wide open joint passages are important in that they allow the walls to be exposed to rain, wind, and sun.

Rock fins are also found as sandstone ridges which separate parallel drainages or two opposing valley headwalls. The fin development accompanies the later stages of entrenchment and valley widening of these stream courses. Salt Canyon in Canyonlands National Park exhibits this type of fin development.

DEVELOPMENT AND GEOMETRY OF ARCH OPENINGS

Many factors interact to produce a resultant arch. These are discussed under the headings, Lithology, Structure, Climate, Weathering and Erosion, and Time.

Lithology

A natural arch will form only in a rock competent enough to support a roof (Blair and others, 1975); therefore, it is not surprising that the most stable arches develop in sandstones. There are a number of sandstone formations in southeastern Utah competent enough to maintain an arch, but, curiously, nearly all the known arches are confined to either the Entrada Sandstone or Cedar Mesa Sandstone. Arches have been noted in the Juassic Navajo Sandstone and Triassic Windgate Sandstone, but they are exceedingly rare.

None of these sandstone formations are truly homogeneous; thus, there are horizons within them which are less competent. It is observed that initial breakthroughs most often occur along horizontal re-entrants composed of calcareous mudstones or shale partings. The re-entrants or hollows develop at the same level on opposing sides of the fin, and as they grow, eventually intersect to form a breakthrough or an incipient arch. In Arches National Park, the arches found in the Windows section develop at the Dewey Bridge Member-Slick Rock Member contact in the Entrada Sandstone (Lohman, 1975).

Some arches, such as those in the Devil's Garden section of Arches National Park show no obvious lithologic contact.

A qualitative observation of the degree of effervescense of dilute HCl applied to the sandstone suggests that the per cent of carbonate cement varies vertically within the Slick Rock Member of the Entrada Sandstone (Blair, 1975). The richer carbonate horizons frequently reveal themselves by leaving a string of small holes or hollows parallel to the bedding. Although not confirmed by petrographic studies, variations in the per cent of clay cement may also lead to those less competent horizons.

Structure

All of the arches observed in Canyonlands and Arches National Park are confined to sandstone layers which are horizontal or gently dipping, such as fold limbs and, therefore, are exposed over a broad region. This partly accounts for the wide distribution of arches in this part of Utah. Vertical joints crosscutting the fins are rare. Where they do occur, arches usually do not form; although in Horse Canyon in Canyonlands National Park, openings have developed along lower portions of joints and still have allowed a small roof (marred by the joint) to exist.

In addition to the master joints which control rock fin formation, tensional unloading joints were found which closely mirror the existing surface morphology. These joints are due to the weight of the supporting rock column and possibly the release of residual or locked in stresses from the time of lithification. These unloading joints are the direct response to the internal distribution of stresses within the rock. Dr. Elsaged Ahmed Eissa (Hoek and Brown, 1980) has modeled the distribution of the principal and minor stresses and the stress trajectories of various tunnel openings within an idealized homogeneous rock (figs 1, 2 and 3). These diagrams tend to support the observations of tensional joints forming in the regions along the upper walls and ceilings where tensional stresses overcome low compressive stresses. Expansion occurs along these joints normal to the rock surface as defined by stress trajectories within the rock (lefthand of figures) and causes exfoliation of rock slabs from the surface.

The removal of these arcuate shell-shaped slabs both shape and enlarge alcove and arch openings (Hunt, 1953; Lohman, 1975; Gregory, 1938). The arch design is the most stable geometric form an opening can take because it distributes most efficiently the release of internal stresses (Blair and others, 1975).

Depending on the distribution of stresses, variations in the arch geometry will ensue. Compare figure 1, 2, and 3. In figure 1 where the horizontal or lateral compressive stress is twice the vertical load, potential tensional failure will occur at the sites where the compressive stress values are least which is at the side walls. The stress trajectories indicate that if lateral expansion takes place, a

more oblong geometry will result. The most extreme oblong arches such as the Landscape Arch in Arches National Park have thin roofs and support little vertical load compared to horizontal load. Figure 2 represents a situation of equal vertical and horizontal compressive stresses. Under these conditions, a more circular shape is favored. Figure 3 models an opening subjected to a large vertical load compared to horizontal compressive stress. In this instance, a vertical egg shape would be most stable as indicated by the stress trajectory pattern. Flat-lying oblong arches tend to be common in southeastern Utah while true circular arches are rare. No true upright egg-shaped arches were found suggesting that horizontal loads dominate the arch formation. Several vertical-oriented arches are known but their shapes are irregular indicating the influence of other factors in their development. Also note that as an arch grows larger, the surface area under an arch will increase by a length factor squared compared to the overlying volume or mass which increases by a length factor cubed. This will constrain the upper size limit of the arch opening. If the upper range of surface area to load limit is exceeded, the roof will collapse.

A uniform arch shape is rarely achieved because neither the arch-forming processes nor the internal stress distribution of the sandstone are homogeneous. Smaller arches tend to have more irregular openings than larger arches because the smaller arches are stable enough not to necessitate the arch shape.

Climate

The arches are concentrated in the semiarid southwestern desert and one is inclined to think that the arid climate is required for their evolution. This may be fortuitous because the rate and intensity of both weathering and erosion processes were probably accelerated during more humid times in the past. This occurred during the waning stages of glacial advances as suggested from the pluvial history of the Great Basin (Morrison, 1965). The existing climate, however, does contribute to the ongoing evolution of arches. The region around southeastern Utah averages about 20 cm/yr precipitation which is distributed as snow and sleet in the winter and through convection storms during late summer. Although the moisture is slight, its presence is enough to fuel weathering and erosion activity.

Weathering and Erosion

The back walls of most alcoves (pre-arch stage) are flaking off with thin (usually less than 1 cm thick) chips which are coated on their back sides with a thin layer of white salt. The dominant salt observed leaching from sandstones and shales is thenardite (Na_2SO_4).

Most likely a combination of hydration, wet-dry cycles, and possibly salt weathering operate simultaneously to loosen the surface veneer. At the back of many alcoves, there is evidence of water having dribbled out of the rock, especially just above a shale or less permeable horizon than the overlying sandstone. Some even support active springs. The presence of this moisture accelerates weathering and, thus, alcove development. Freeze and thaw also operates but its specific impact is not known.

The removal of surface materials is by gravity, wind, and water. Wind is important in arch formation only as an agent which removes loose grains from the surface. No sandblasting effects were observed.

Time

The rate of arch formation can be rapid or slow depending upon the influence of many variables. If exfoliation fractures develop, the opening may form rapidly from the collapse of slabs as occurred with the Skyline Arch in Arches National Park in 1940. Most arches continue to enlarge slowly as grains of sand are gently loosened and removed. These arches exhibit smooth, rounded surfaces in contrast to arches with angular edges associated with fracture release (Blair and others, 1975).

No criteria was observed for assigning absolute ages to the arches. Desert varnish has been found on the exterior of many of the canyon walls but has not been dated in this area of the Colorado Plateau.

CONCLUSION

The evolution of an arch is dynamic and complex. It involves the delicate interplay between rock, structure, process, and time. The magnitude of input from each of these factors determines the style of arch formed. One significant realization from this study is that each arch is a separate entity and that the factors controlling the inception and growth differ from one arch to another. Hence, there exists a wide variety of arch shapes and sizes.

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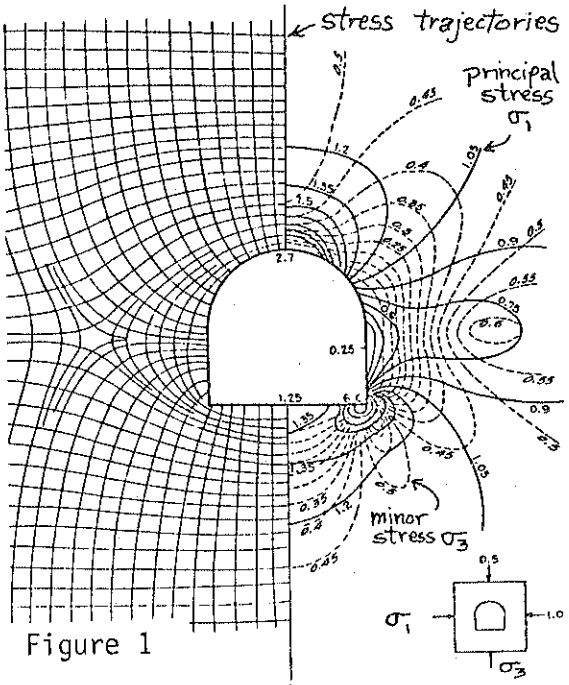

Figure 1

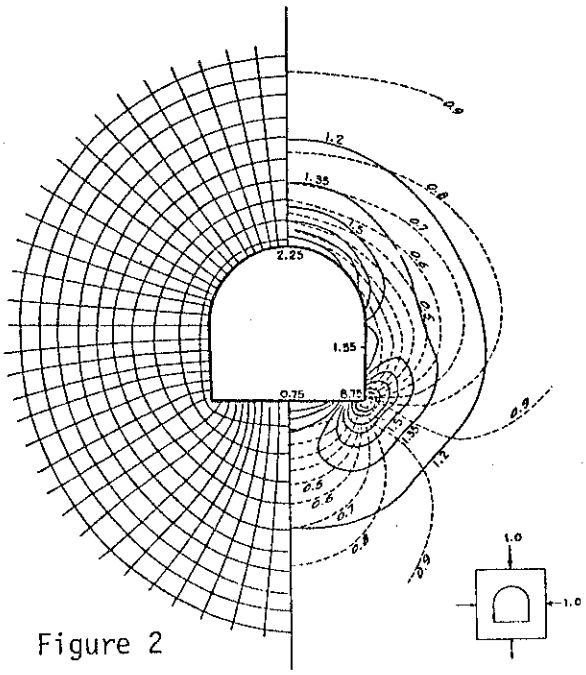

Figure 2

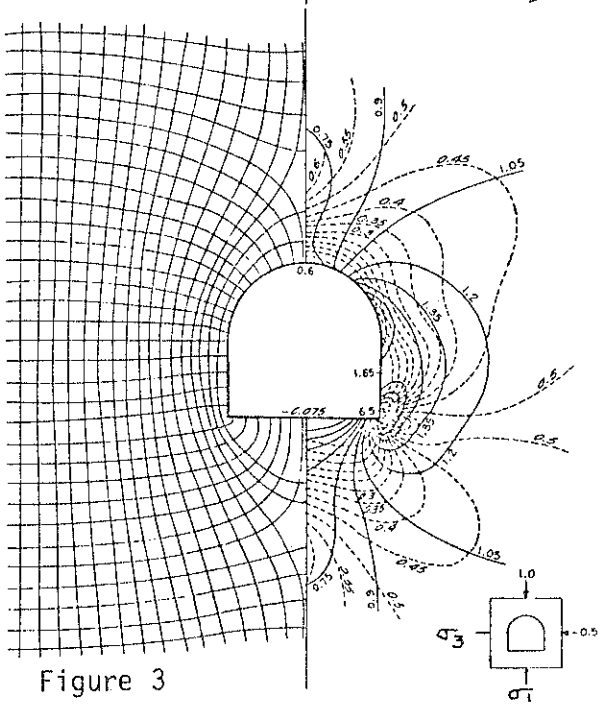

Figure 3

Figures modified from Hoek and Brown (1980)

THE LATE PALAEOZOIC GLACIATION OF SOUTHERN SOUTH AUSTRALIA

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Abstract

This paper discusses evidence for Late Palaeozoic glaciation in southern South Australia and considers various controversial issues related to it. These include factors concerned with the nature of the glaciation, the direction of ice movement, the precise age of the glaciation, the question of multiple glaciation and the possibility of glacial land forms persisting in the modern landscape.

The Late Palaeozoic ice mass was of continental proportions and wet-based and temperate in character. It passed across southern South Australia in a general southeast-northwest direction, but superimposed upon this flow were numerous local variations in ice motion. The precise age of the glaciation has still not been established and it may straddle the Carboniferous/Permian boundary. Contrary to work elsewhere in Australia, there is as yet no definitive evidence supporting multiple Late Palaeozoic glaciations in southern South Australia. Although some glacial landforms, preserved by burial, persist in the modern landscape, some others ascribed a glacial genesis, may be of non-glacial origin and of post-Permian age.

Introduction

The passage of glacier ice over the South Australian landscape has long been known; in fact, the first discovery of glacial action in Australia was made in the Inman Valley, south of Adelaide, by A.R.C. Selwyn in 1859 (Selwyn, 1860). Interest in the imprints of glacial activity in South Australia was reawakened by the discovery of further evidences of glaciation at Hallett Cove (Tate, 1877;1884), although some scepticism greeted this pronouncement. In particular, Scoular (1884) preferred to explain the striated bedrock in terms of water erosion and sand driven by the north wind. David and Howchin (1897) made further finds related to the evidence of glaciation when they sought the original Selwyn Rock, and in the ensuing 30 years or so, Howchin published prolifically on the glaciation, demonstrating its occurrence on Fleurieu Peninsula, Kangaroo Island and Yorke Peninsula. He must rank as the pre-eminent and most perspicaceous worker on this aspect of South Australian geology and geomorphology, and the majority of his general conclusions regarding the glaciation are valid today [see Royal Society of S. Aust., Transactions 1933 (p242-29) for a bibliography of Howchin's works up until 1933.]

Subsequently there have been many controversies surrounding the glaciation in South Australia, particularly in regard to the nature of the glaciation, the direction of ice movement, the age of the glac-

iation, the number of glacial episodes and the possible preservation of landforms related to the glaciation.

Evidence for the Glaciation

Indisputable evidence for an extensive glaciation in southern South Australia is provided by numerous striated surfaces, the occurrence of erratics from both local and exotic sources, a wide range of distinctive glacigene sediments, including lodgement and ablation tills, ice contact fluvioglacial sediments, outwash fan deposits and glacio-lacustrine and glacio-marine deposits with associated dropstones. The distribution of glacigene sediments of Late Palaeozoic age is shown in Fig. 1. Those in the north of the state are of particular interest because they are associated with commercial gas and oil deposits.

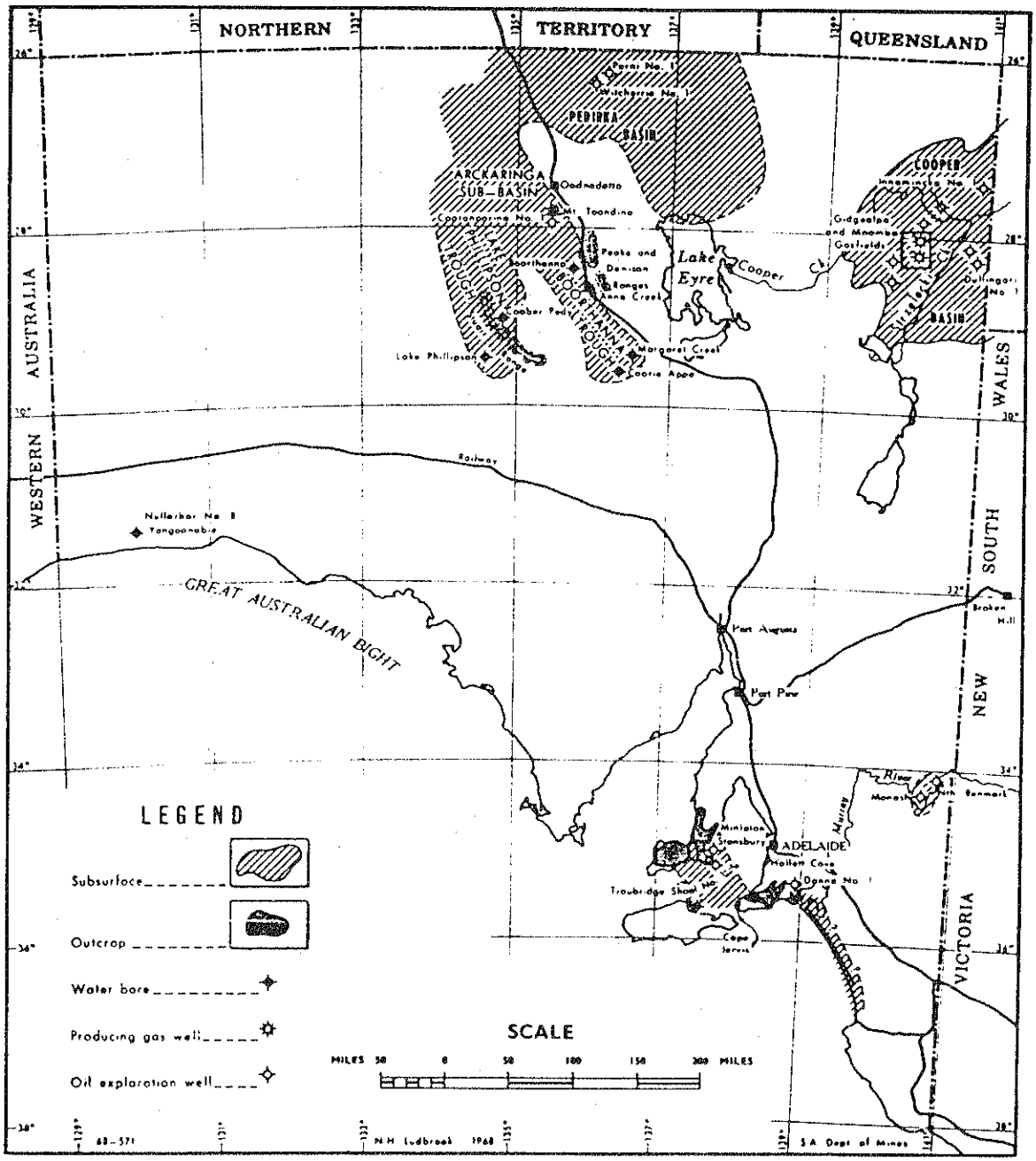

Figure 1 Areas of Late Palaeozoic sedimentation in South Australia (from Ludbrook, 1969)

Age of the Glaciation

Selwyn (1860) made no assertion concerning the age of the glaciation, but Brown (1892) associated the striated Selwyn Rock with the adjoining glacigene deposits, which he mapped as of Tertiary age. Tate (1877) associated the glaciated bedrock at Hallett Cove with the Pleistocene glaciations of the Northern Hemisphere, but excavations later demonstrated its pre-Pliocene age (Tate, Howchin and David, 1895). Howchin (1926) correlated the glacial deposits with the fossiliferous glacigenic sediments of Bacchus Marsh and thus assigned them a Permo-Carboniferous age. On the basis of some apparently dubious evidence, Segnit (1940) suggested a Lower Cretaceous age for the glacial beds at Hallett Cove, but this was refuted by Mawson (1940) and Sprigg (1942).

Prior to the work of Ludbrook (1957;1967) no diagnostic fossils had been found in the glacigene sediments. Her discovery of brackish water, arenaceous foraminifera of Lower Permian (Sakmarian) age in the glaciegene sediments apparently fixed the age of the glaciation, particularly as she considered that the fossils came from near the base of the sedimentary sequence. However, Alley and Bourman (1984) showed that the marine fossiliferous unit represents only the final deglacial stage of the glacial episode at Cape Jervis, so that the possibility still exists that the glaciation straddled the Permo-Carboniferous boundary. This question remains to be resolved.

Nature of the Glaciation

Two opposed views have been presented concerning the nature of the glaciation; Campana and Wilson (1955) favoured a mountain or valley type of glaciation, whereas Howchin (1926) envisaged a terrestrial glaciation of continental proportions.

Campana and Wilson (1955) maintained that the over-deepening of valleys and the presence of glacial bars and cirques in the area are indicative of a valley type glaciation. The amount of valley deepening is considerable; for example a bore at Back Valley revealed that the bedrock glacial floor is some 240 m below sea level, while glacial sediments occur up to 300 m above sea level to give a total surviving glacial relief of some 540 m. However, glacial deepening and irregular bedrock topography are not restricted to valley glaciations. Moreover, the recognition of alleged cirque forms as evidence of a mountain type glaciation is questionable, particularly in view of reconstruction of the glacial relief and the evidence of extensive erosion in post-glacial times.

If the site of the present Mount Lofty Ranges had been the source of the ice mass, and there is no evidence of any other nearby range, then there should be evidence of glacial movement radiating from this area, but all of the evidence suggests a movement of ice from the southeast, overriding the remnants of the dissected fold mountain range.

Howchin (1926) reconstructed evidence of ice movement from Tasmania, Victoria and South Australia and concluded that ice spread from a central point located to the southwest of Tasmania and the glaciation was thus of continental extent and essentially terrestrial in character.

The ubiquity of marine arenaceous foraminifera in the main sedimentary basins, apart from a section of the Great Australian Basin, led Ludbrook (1967) to suggest that there had been considerable deposition in a fiord-like marine environment. However, Alley and Bourman (1984) have shown that the marine environment only persisted in the final stages of deglaciation, and that the sedimentary sequence and the erosional forms indicate the passage of a wet-based temperate terrestrial ice mass in the early stages of the glaciation.

Thus, at its maximum extent, the ice mass appears to have been a large scale terrestrial glacier. However, both during the onset of glacial conditions and during deglaciation, discrete ice tongues and cirque glaciers may have existed, giving rise to forms akin to those of valley glaciers, while in the final stages, particularly in coastal lowland environments, fine-grained sediments were deposited in sea water diluted by abundant meltwaters.

Insofar as the extensive post-glacial erosion of the Mount Lofty Ranges allows, it appears that the ice mass at its maximum was at least 1000 m thick and the topographic relief of the ranges was possibly double this value.

Direction of Glacial Movement

Evidence for the direction of glacial movement has been derived from three main sources; striated rock surfaces, the provenance of erratics and fabrics of lodgement tills.

Today there are some 25 separate areas of glacially striated rock surfaces known on Fluerieu Peninsula, Kangaroo Island and at Hallett Cove. These surfaces variously display grooves, striae, polish, various forms of friction cracks, 'p' forms and micro-erosional features. Care needs to be taken to distinguish genuine glacial striae from similar markings resulting from tectonic activity and mass movements.

Some erratics are particularly useful as indicators of ice movement, but many of the larger erratics appear to be ice rafted so that they may not be as critical in this regard.

Where neither striated surfaces nor indicator erratics occur, fabrics of basal lodgement tills have occasionally been utilised to throw light on the direction of ice travel. Using a combination of these various criteria, some reasonably useful reconstructions of ice flow have been made. Figure 2 illustrates the direction of ice movement across Fluerieu Peninsula from exposures of striated bedrock.

Several earlier workers (eg Crowell and Frakes, 1971a, 1971b) considered that the direction of ice movement across South Australia was generally from the south to the north. However, distinctive erratics derived from igneous rocks in the Murray Basin and from Western Victoria suggest that the ice passed across southern South Australia from the southeast to the northwest. Moreover, superimposed upon this general direction of travel were various deviations as revealed by striae and till fabrics. For example, when the Mount Lofty Ranges were reached it appears that the ice was funnelled in a more westerly direction through the Inman Valley across Fluerieu Peninsula (Milnes & Bourman, 1972; Bourman & Milnes, 1976 and Bourman et al. 1976) and through Backstairs Passage between Kangaroo Island and the mainland. On the western side of the ranges the ice met another lobe moving in a more northerly direction along the site of the modern Gulf St. Vincent and continued across the gulf to Yorke Peninsula after reverting to the more northwesterly flow.

In places the ice flow had been apparently affected by bedrock topography such as at Hallett Cove where a bedrock depression caused the ice flow to vary from 280° to 15° (Sprigg, 1945; Milnes and Bourman, 1972). However, at Nunn Valley, on central Fleurieu Peninsula, ice passed transversely across a pronounced structural valley developed along the contact of Archaean and Proterozoic bedrock. In detail, much variability in striae direction is revealed on small stoss and lee features. At Blinman in the Flinders Ranges, an isolated outlier of basal glacigene sediments indicates a general northerly flow of ice.

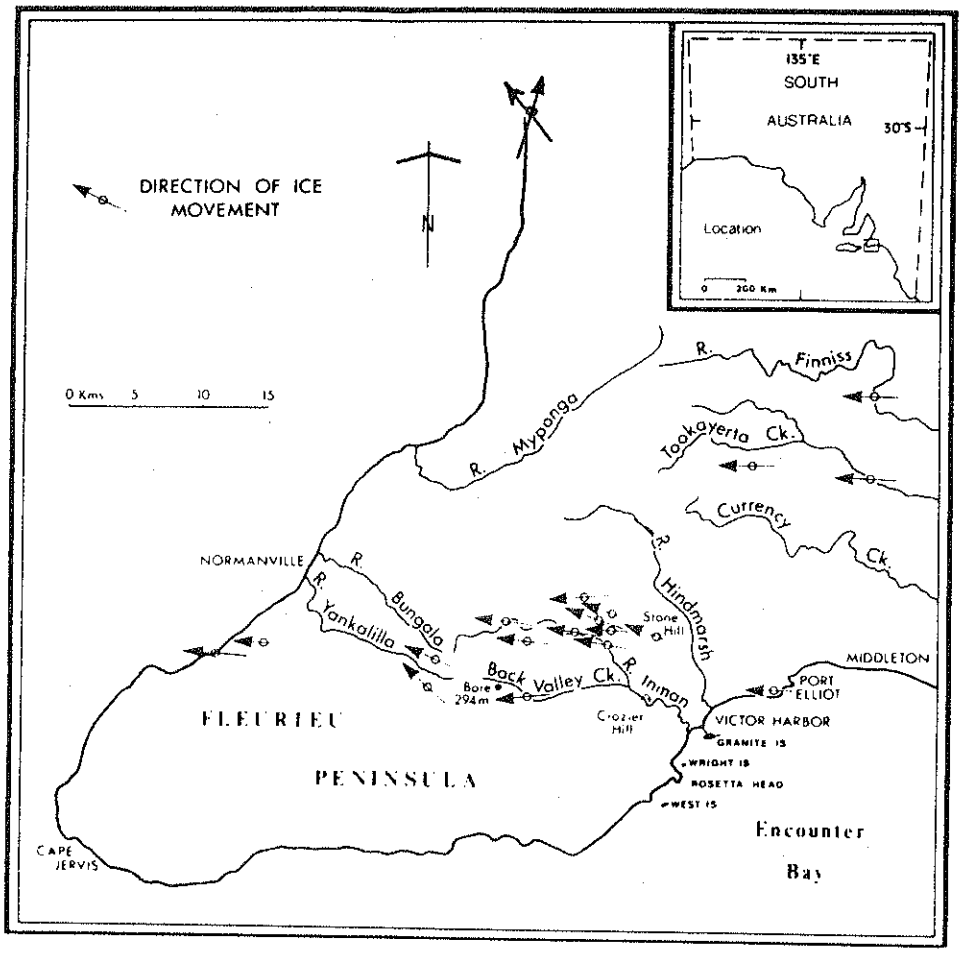

Figure 2 Striated rock surfaces, Fleurieu Peninsula. (Various sources)

Multiple Glaciation Evidence from Southern South Australia

Various workers have proposed multiple phases of glaciation in southern South Australia, but to date no crucial evidence suggesting that any more than one glaciation occurred has come to light.

Crowell and Frakes (1971a, 1971b) considered that crossing striae at Hallett Cove indicated readvance of a second ice mass across a previously striated surface. However, the several sets of striae do not bear consistent cross-cutting relationships and can be adequately explained by alteration in ice movement during one glacial advance by variations in ice thickness.

Maud (Unpub MS) suggested that lithified fluvioglacial sediments, squeezed sediments and drumlin-like features represented evidence of multiple glaciation, but all of the features are equally explicable in terms of fluctuating ice mass during a single glaciation.

Bowen (1958) interpreted deposits at Cape Jervis and Hallett Cove as containing multiple tills that indicated two separate glacial advances at the former locality and three at the latter. Recent work at Cape Jervis (Alley and Bourman, 1984) demonstrates that there is only one genuine lodgement till at the base of the sequence and that other till-like materials are actually flow tills that derived by sliding from the ice surface and from icebergs

during the decay and stagnation of the ice mass. The section at Cape Jervis appears to reveal evidence of the deteriorating conditions prior to the glaciation, the passage of a wet-based terrestrial ice mass and successive environments that reflect the progressive decay of that ice mass. Work at Hallett Cove (Bourman and Alley, in prep.) reveals a similar sequence, with only one lodgement till being present, but with numerous diamictons(up to nine) that superficially resemble lodgement tills, but which we regard as flow tills developed during deglaciation.

Nowhere in southern South Australia have unquestionable interglacial sediments been reported to separate genuine glacial deposits, nor has more than one complete glacial and deglacial sequence been described. Consequently, there appears to have been no more than one glacial episode related to the Late Palaeozoic glaciation preserved in southern South Australia.

Glacial Landforms

Numerous glacial landforms have been recognised and described in southern South Australia by various workers, and these include roches moutonnées, U-shaped valleys, hanging valleys, a crag-and-tail, perched blocks, cirques, kettle holes and drumlins. Some glacial landforms occur essentially unmodified in the modern landscape due to prolonged burial by glacigene sediments and only recent exhumation. However, great care is required in ascribing glacial origins to particular landforms because some alleged features have developed via non-glacial processes in post-Permian times, and some Permian features have been so severely eroded that it is impossible to reconstruct their forms with any degree of certainty.

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SOIL DEVELOPMENT IN THE BARRIER RANGE, WESTERN NEW SOUTH WALES

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ABSTRACT

Soil development in the Barrier Range reflects not only a combination of the currently arid climatic regime and underlying soil parent materials, but also the legacy of Late Pleistocene and Holocene palaeoenvironmental conditions. At present mean annual rainfall in the area is approximately 200mm, although the amount is highly variable from year to year. Potential evapotranspiration rates greatly exceed precipitation, even in relatively wet years. Under such climatic conditions rates of soil formation can be expected to be very slow, particularly on many of the hard quartzites, dolomites and sandstones of the hilly country. Even on softer shales, silstones and colluvial/alluvial deposits soil forming processes are generally limited to the minor redistribution of some of the soluble salts and the development of pedological structures in the upper few centimetres of profile. In reality, however, much of the range and adjacent flat country is mantled by relatively deep (1.0 - 1.5m), clayey soils. It is also of common occurrence to find shallow (0.5m) red clayey soils overlying unweathered quartzites on ridge crests.

Detailed chemical, physical, mineralogical and micromorphological studies of the red clayey soils of this mantle (Chartres, 1982; 1983a; 1983b), suggests that much of the soil parent material was deposited by aeolian activity during the last major arid phase of the Late Pleistocene period. The soils developed in the mantle usually have texture contrast profiles (Dr 1.13; Northcote Key) and are known as desert loams in the Handbook of Australian Soils (Stace et al., 1968). According to the Soil Taxonomy (Soil Survey Staff, 1975) they key out as Typic Natrargids. In places the loamier textured A horizons are absent, in which case the soils key out in the Uf group of Northcote, and in some areas they overlie silica hardpan, which places them in the Dr 1.16 group of Northcote and makes them Typic Nadurargids in the Soil Taxonomy. Characteristically, desert loam soils have thin (0 - 5cm) loamy A horizons and stone-mantled surfaces. Some of the A horizon material is colluvial in origin and some of the stones have also been derived from upslope positions under the influence of associated gravitational processes. However, it is also considered probable that many of the larger stones were kept at or near the soil surface during the phase of aeolian deposition due to a combination of rain-wash and shrink-swell processes. The B horizons are virtually stone-free, highly pedal, sodic and sometimes over 1m deep, with clear boundaries to the underlying Bcs and C horizons. The Bcs and C horizons are marked by accumulations of calcium carbonate, soluble salts and gypsum, and are often stony. It is considered that they are probably the truncated remnants of older soil profiles, which have been buried by aeolian silts and clays, which comprise the current B horizons.

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Where aeolian deposits do not occur, usually because of subsequent erosion, a range of other soil types are found. On shales and siltstones, highly alkaline, solonized brown soils (Gc 1.1 and 1.2 to Uc 1.11 and 1.12; Typic Calciorthids) are common. On steep rocky slopes lithosols are found, whereas siliceous sands are found interspersed with cracking clay soils on some of the adjacent plains.

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DEEP LEADS AND LANDSCAPE INTERPRETATION IN NEW ENGLAND, N.S.W., WITH VICTORIAN COMPARISONS

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ABSTRACT

Deep leads, or buried fluvial sediments, were investigated in the Armidale area of New England as early as 1852, and extensively mined for gold. Records of this early work, in conjunction with more recent bore logs, has enabled the deep leads to be mapped. They trace out two drainage networks of quite different ages. The older network, absolute age unknown, lies low in the modern landscape, and is flanked by younger, dendritic leads. These younger leads mark the course of a north-trending stream system that was blocked and reversed by Oligocene basalt flows.

In comparison with sub-basaltic deep leads in New England, those in Victoria span a greater relief range, and are much younger, having been buried under Pliocene-Pleistocene basalts. The sediments in the Victorian deep leads were deposited under a drier, cooler climate, marked by alternating glacial and interglacial periods.

Sub-basaltic deep leads contain economic concentrations of gold, tin and gems. Besides this, they aid landscape interpretation; they delineate relict stream courses; aid in the reconstruction of pre-basalt landscapes; preserve a record of past depositional environments; and serve as a datum point for calculating post-basaltic valley incision.

1. INTRODUCTION

In mining terminology a lead refers to the presence of geological conditions likely to result in mineral deposits. Deep leads are so called because of their sub-surface occurrence. In Australia the term usually refers to buried fluvial sediments.

I studied the distribution of deep lead systems during research into landscape evolution in the Armidale area, New South Wales (Connolly, 1983), (Map 1). Discussion of these leads, and comparison with Victorian examples, form the subject of this summary paper.

2. HISTORY OF DEEP LEADS IN THE ARMIDALE AREA

Investigation of deep leads in the Armidale area followed the discovery, in the early 1850's, of gold in modern stream sediments. The first deep lead gold was discovered in 1852 when

gold-bearing sand was traced to its source beneath a Tertiary basalt capping on a hill 3 km west of Uralla (David, 1886).

Following this, there appears to have been a growing realisation that sub-basaltic deep leads in the area between Uralla and Armidale were likely sources of gold. As a result exploration of the deep leads became the focus of miners' attention.

By the 1880s the extent of the richer sub-basaltic gold deposits between Uralla and Armidale had been determined from scores of shafts sunk by hand through basalt up to 57m thick. Extensive tunnelling also took place. For example in 1877 The Long Tunnel Company attempted to tunnel between Sydney Flat and Doherty's Hill, in an attempt to drain the water-saturated deep lead that had been found under Doherty's Hill. The tunnel was 'pushed forward for a distance of about 1,800 feet' before lack of funds forced the Company to cease operations (David, 1886).

Mining of gold from deep leads in the area continued until the 1950s.

3. MAPPING DEEP LEADS IN THE ARMIDALE AREA

Since these early explorations, there has been little direct investigation of deep leads in the area between Uralla and Armidale. However, there have been many water bores sunk in the area, and these are the main source of new information about sub-surface geology.

From this historical and contemporary sub-surface information, and on the basis of detailed mapping of surface materials and surficial geology, it is possible to draw cross-sections showing sub-surface geology. From these cross-sections the location and extent of deep leads in the area between Uralla and Armidale can be fixed. Once this is done the surface distribution of the deep leads can be mapped. (Map 2).

4. DEEP LEADS AND LANDSCAPE INTERPRETATION

Several points are evident once the deep leads are mapped.

1) The deep leads invariably underlie basalt, though some are now partly re-exposed by erosion.

2) The deep leads are not necessarily low in the modern landscape. Relief inversion has operated widely through the New England Region, and basalt remnants and the leads underlying them are often topographic highs.

3) Deep leads in the area between Uralla and Armidale trace out two drainage networks of different ages. The oldest, trending north-south through the Saumarez-Arding area, is an area of low relief in the modern landscape, approximately parallel to, and 0.5-1.0km east of, the Main Divide. Immediately east and west of this older basalt-covered deep lead, are lateral deep leads underlying basalts that form topographic highs. The western lead underlies the basalts that cap the Main Divide; the eastern lead underlies the basalts of the divide between Saumarez Creek and Dumaresq Creek.

   The lateral deep leads trace out a dendritic north-trending stream system. This stream system is flowing in the opposite

direction to the modern drainage, and indicates that drainage reversal took place concurrently with the Tertiary basalt extrusions.

4) The possible causes of drainage reversal in the area are either localised faulting or uplift, or the damming of pre-basalt valleys by lava flows. There is no field evidence for localised faulting or upwarping of the sub-basalt surfaces, but there is a marked increase in the thickness and extent of basalt remnants to the north of Armidale. So the most likely cause of the drainage reversal to a south-flowing drainage system, is the piling-up of basalts to the north of Armidale, close to possible volcanic vents in the Guyra region.

5) Silicification of deep lead sediments is common, particularly towards the lateral margins of leads. This suggests variations in the silica content and flow characteristics of groundwater during the period of silcrete formation.

5. COMPARISON WITH DEEP LEADS IN VICTORIA

Deep leads in Victoria, as in New England, are commonly sub-basaltic, but they span a far greater relief range. In north central Victoria, leads vary from 95m below to 125m above the modern landsurface (Williams, 1983). These Victorian deep leads, like those in New England, have a history of intensive investigation, mainly because of the gold contained in them.

The sub-basaltic Victorian leads have been preserved under basalts of Pliocene to Pleistocene age, and are thus much younger than their Armidale counterparts. As noted by Williams (1983), sections of pre-basalt streams in Victoria that were never filled by basalt are often preserved beneath thick layers of clay and silt.

The age difference between sub-basaltic deep leads in Victoria and New England is important because it means that the deep lead sediments were initially eroded, and transported within a fluvial depositional environment, under quite different climatic regimes.

The climate during the period of Oligocene volcanism in New England was warmer that the present, despite marked cooling and a decrease in mean global precipitation at the end of the Eocene (Frakes, 1979). The build-up of the Antartic ice sheets during this time (Frakes, 1979) suggests that there was abundant moisture moving across Australia. In summary, the pre-Oligocene basalt sediments were weathered and transported under a climate that was probably substantially warmer and more humid than the present.

The Pliocene-Pleistocene epochs were times of widely fluctuating climate, with the onset of alternating glacial-interglacial intervals. Mean global precipitation fluctuated with the temperature variations (Frakes, 1979).

These climatic changes mean that although New England and Victoria both experienced a prolonged period of deep weathering during the Mesozoic and early Tertiary, the Victorian deep leads may preferentially contain sediments liberated towards the end of this prolonged weathering phase. In Victoria, much of the pre-Newer Basalt sediments, including an unknown fraction of gold, were probably transported right through the river systems prior to the valleys being infilled by basalt.

6. CONCLUSIONS

Deep leads in the Armidale area have been of great economic importance, and are still the subject of investigation by mining companies for the gold and other minerals they contain. The same is true of deep leads in Victoria. Apart from this economic significance, deep leads are of considerable value in landscape interpretation for the following reasons:

1) They mark the course of relict stream systems, and in the absence of localised warping, also help in the working out of pre-basalt topography.

2) Where deep leads underlie dated basalts they are a record of the depositional regime at the time of volcanism, and enable comparison of variations in depositional environments in different places and through time.

3) The topographic position of deep lead sediments gives a datum point from which post-basaltic valley downcutting can be calculated.

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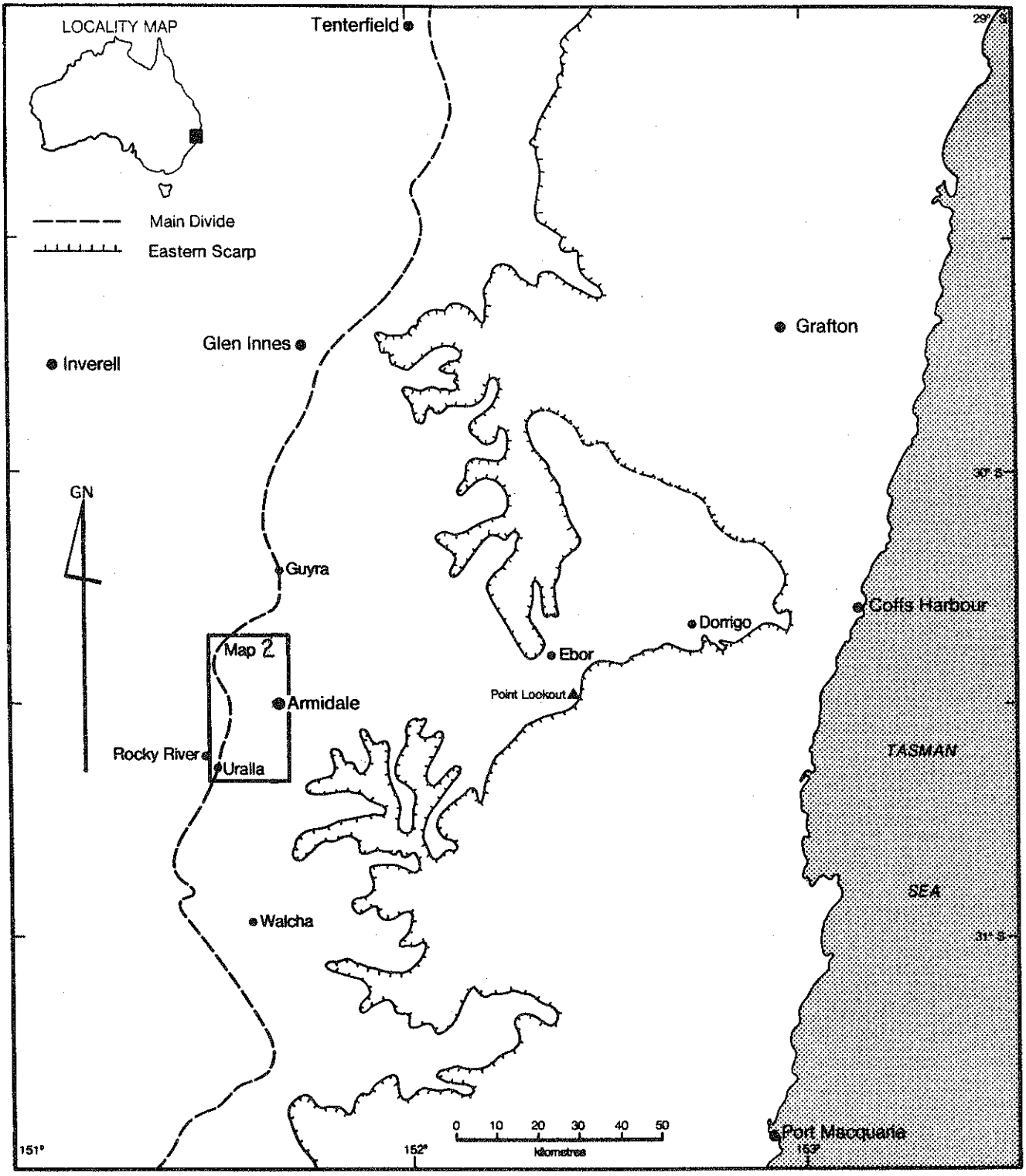
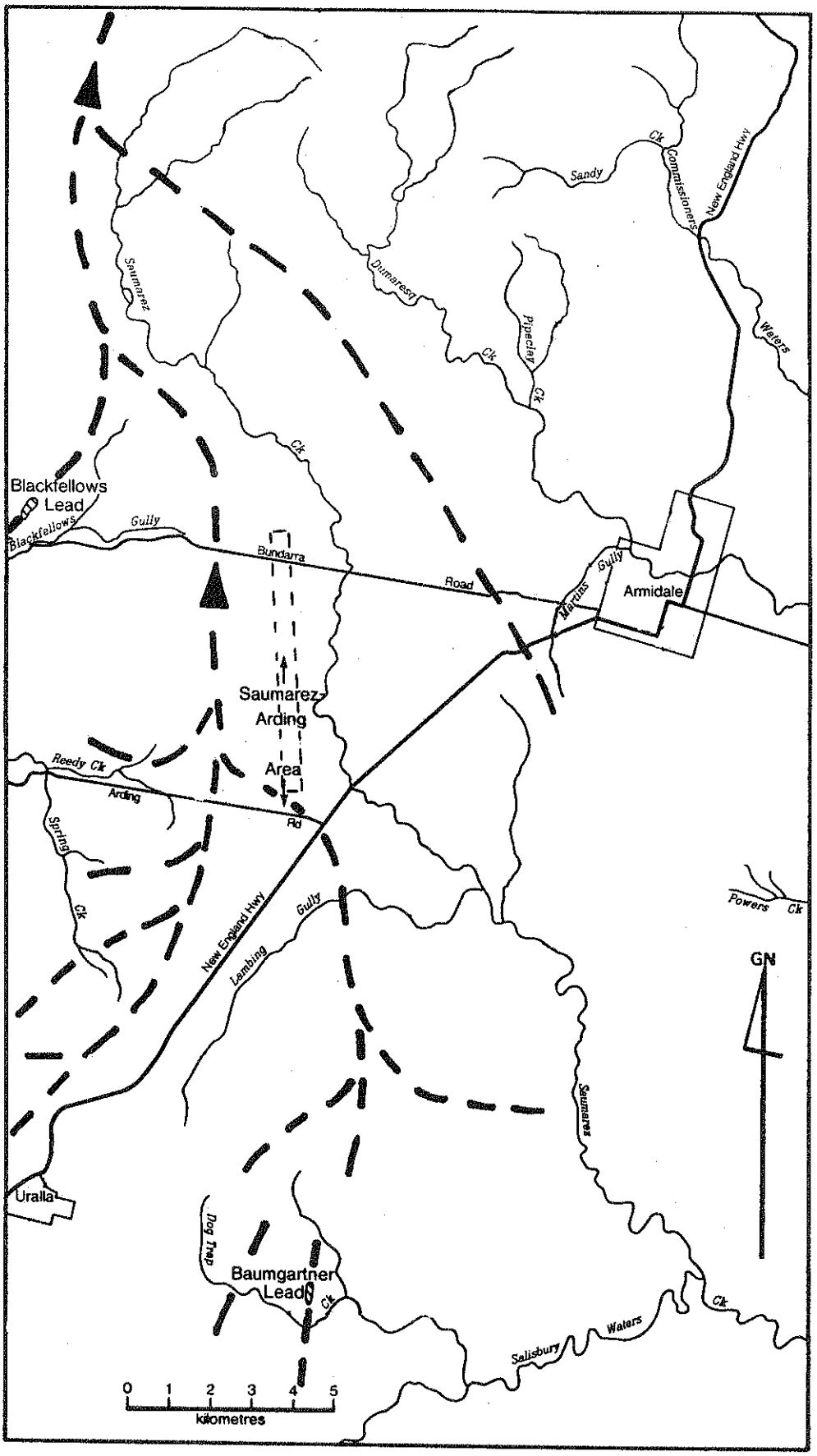


Map 1. Location of the Armidale study area.



Map 2. Deep leads in the Armidale area. Light dashed lines show trend of older deep lead. Heavy dashed lines show younger, lateral leads.

SOIL DISTRIBUTION ON AEOLIAN LANDSCAPES IN SOUTH-WESTERN NEW SOUTH WALES

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ABSTRACT

Aeolian landscapes of far south-western N.S.W. have been mapped into soil-landform associations to test susceptibility to wind erosion. The soil groups encountered are described and hypotheses on their formation are offered.

INTRODUCTION

The far south-western corner of New South Wales is dominated by Quaternary aeolian landscapes. These relict landscapes have developed from water deposited sediments (Pels, 1969) although the widespread occurrence of highly calcareous material suggests that wind borne clay (parna) may also have contributed sediments to the area (Butler, 1956). The area is characterised by sandplains, often with superimposed dunes, lunettes, hummocks and ridges, with alluvial intrusions.

The surface soils are geologically recent, having developed from weathering and reworking of the underlying source materials. The calcareous soils of the plains and linear dunes have developed from calcareous clayey sands and the parabolic dunefields from loose, siliceous sands. Contemporary soil distribution and landform patterns are strongly related, as are soil physical and chemical properties.

The generally coarse texture of the soils makes the landscape inherently susceptible to wind erosion when vegetative protection is removed. Clearing and cultivation have increased rapidly over the past decade (Alchin 1980) and have resulted in a number of problems in the area. These have been discussed by Eldridge and Semple (1982) and Lievers and Luke (1980).

In order to determine land management strategies for the area, a project has been initiated to define soil-landscape types and to subject the major ones to simulated wind erosion using a wind tunnel.

The Study Area

The study area is bounded by 141° and 146° E, and 33°S and the Victorian Border. Aerial photography interpretation, field mapping and soil sampling have been carried out on the major aeolian landforms and soils. The Riverine Plain has been excluded because

the predominantly fine-textured soils are not subject to widespread wind erosion. Soils have been classified according to their principal profile forms (Northcote 1971). Maps (published separately) have been drawn at a scale of 1:100 000. The location of the survey is shown in figure 1.

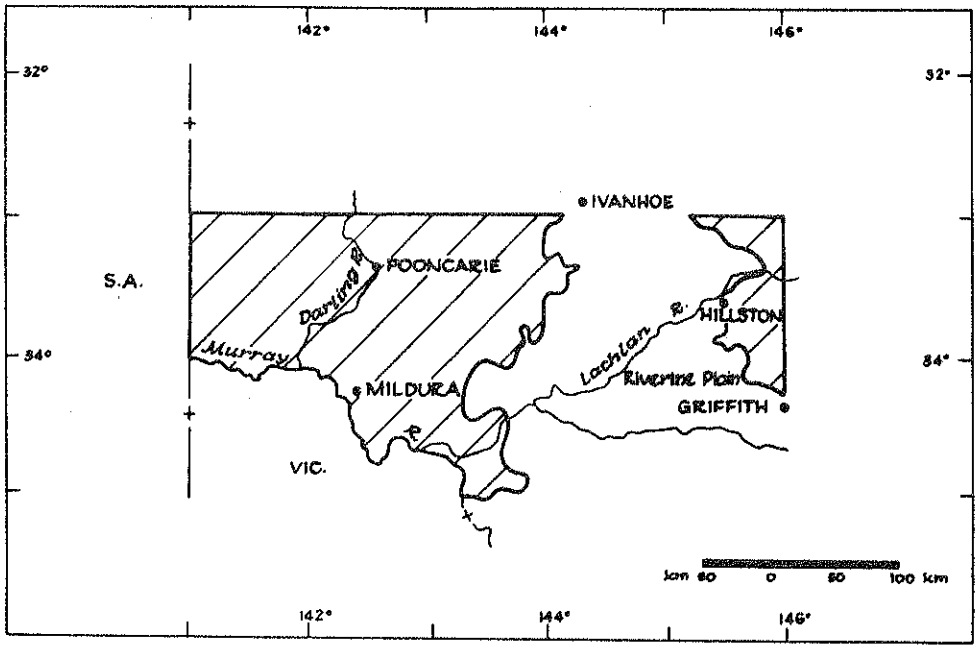

Figure 1. Extent of Aeolian Landscapes in Far South-western N.S.W.

Very little detailed information is available on soils and their distribution in far south-western New South Wales, except for the Irrigation Areas near the Murray River [Marshall and Walkley, (1937), Northcote (1951)]. Some unpublished erosion surveys have been carried out by the Soil Conservation Service of New South Wales (e.g. Beadle, undated; Jones, 1964; Stannard, 1955) but soil details are limited.

The current survey has been the most detailed to date in the area.

SOIL MORPHOLOGY AND DISTRIBUTION

A total of 33 soil-landform associations have been mapped in the study area. These associations are based on combinations of landform elements and soils, with native vegetation type also usually associated.

Each association comprises several principal profile forms (Northcote 1971), determined by spatial variation and by differences in soil topsoil texture and depth, depth to carbonate, and pedological development, and, less importantly, soil colour and structure.

These differences are due to variations in parent material and to considerable modification of parent material and soils by aeolian activity. As a result, some associations contain the full range of surface textures and principal profile forms across their landform elements.

For the purposes of this outline, the dominant soils have been grouped and described below.

SANDS

Calcareous sands (Uc 1.1) and siliceous sands (Uc 1.2) are typically deep, pale coloured slightly calcareous to non-calcareous sands occurring on sub-parabolic dune formations east and west of the Darling River. These dunes are usually quite high (up to 20 metres relief), sharp crested, and aligned towards the east, and are separated by narrow swales. Little is known of why dunes assume a parabolic rather than a linear form, however Lawrence (1980) proposed that they have developed by having their arms anchored by vegetation while the centre advanced.

Soil formation has been limited by the instability of the material. They occur in association with small areas of pedologically more developed deep brownish sands (Uc 5.11, Uc 5.12) and calcareous earth (Gc 1.12). These soils are formed on material which appears to be similar to the Lowan Sand formation (Lawrence 1975), which is highly leached and non-calcareous.

Brownish sands (Uc 5.11) occur over extensive areas to the east of the Darling River, between "Arumpo" and the Murray River. They are red to reddish-brown at the surface, become calcareous at depth and often contain an impermeable hardpan at depth. Aeolian action has resulted in formation of extensive linear dunefields from sediments similar to the Woorinen Formation (Lawrence 1975), which differs from the Lowan Sands in having higher carbonate and clay contents, which increase with depth. (Neither of these formations has been positively identified in New South Wales).

Within the dune is usually a core of consolidated material of finer texture often with high carbonate levels.

The dunes are thought to have developed parallel to the resultant wind direction by erosion of the adjacent plain (contemporary swales).

The brownish sands occur in association with calcareous sands (Uc 1.13) on dune crests and loamy calcareous earths (Gc 1.12, Gc 1.21) in swales.

The sands normally support mallee (*Eucalyptus* spp.) or belah (*Casuarina cristata*) - rosewood (*Heterodendrum oleifolium*) vegetation types.

CLAYS

Grey cracking clays (Ug 5.2), although essentially lacustrine or riverine in origin, are a characteristic feature of the aeolian landscape. They occur on floodplains, levees, playas, lakes, swamps and terminal drainage basins of the Darling and Murray Rivers, the Anabranch and Willandra Creek. The soils are typically deep, cracking grey to yellow-grey clays with a self-mulching surface and little change with depth apart from an increase in carbonate content.

They occur in association with a range of duplex soils (commonly Dr 2.13, Dr 4.13, Dy 2,13, Dy 2.53 and Dy 4.13) which form scalded levees at the plain margins, and small islands of brownish sands (Uc 5.11) or red massive earths (Gn 2.13). The clays support chenopod shrubs or black box (Eucalyptus largiflorens) communities.

RED EARTHS

Extensive areas of calcareous red earths (Gn 2.13) occur in the eastern section of the study area around Hillston. They are red, massive, loamy soils with small to medium amounts of carbonate in the lower layers. They often have a red-brown hardpan in their sub-surface layers. They occur on extensive level to very slightly undulating plains, sometimes associated with calcareous earths (Gc 1.12, Gc 1.22) or deep brownish dune sands (Uc 5.11), and becoming more calcareous westward.

This group of soils is separated from the main body of aeolian soils further west by the broad expanse of the Riverine Plain. Their origin appears to be in alluvial material from the nearby Lachlan and Warranary Ranges with aeolian additions, resorting and redistribution.

They support mallee or belah-rosewood vegetation.

CALCAREOUS EARTHS

Calcareous earths (Gc 1.12, Gc 1.22) are the most widespread soils in the region. They occur on plains, dunes and low rises on marine sediments modified by aeolian activity. They are calcareous throughout, with clay and carbonate contents increasing with depth. Surface textures range from loamy sands on dunes to clay loams in swales and dune corridors. Often carbonate is exposed at the surface and an impermeable calcrete or kunkar layer may be present at depth.

They support mainly belah-rosewood vegetation with some areas of mallee and small areas of bluebush (Maireana pyramidata).

DUPLEX SOILS

Hard red duplex soils (Dr 2.13, Dr 2.53) occur on plains and along drainage tracts in the east of the study area. They have brown sandy loam to loamy surfaces which set hard when dry. Subsoils are red to reddish-brown highly pedal clays.

These soils occur in association with yellow, crusty duplex soils (Dy 1.33), apedal yellow duplex soils (Dy 4.53), grey-brown calcareous earths, rises of deep siliceous sands (Uc 1.22) and depressions with grey cracking clays (Ug 5.24, Ug 5.25).

They support belah or bimble box (Eucalyptus populnea) communities.

Sandy red duplex soils (Dr 4.13, Dr 4.53) have brown, loose sandy soils overlying red to red-brown clay subsoils. These often have large amounts of calcium carbonate in the B horizon and have been referred to as "solonised red brown earths" by Condon (1961). These soils occur on the plains in association with brown calcareous earths, rises and dunes of deep siliceous (Uc 1.22) or earthy (Uc 5.31) sands and low-lying depressions of grey cracking clays (Ug 5.2) and in some dunefield swales.

The sandy duplex soils have developed in the area transitional between the aeolian plains and the Riverine Plain in the east of the study area. These areas support belah and/or bluebush communities.

All duplex soils are considered to be formed mainly by soil surface redistribution or successive deposition, rather than by clay illuviation. The coarser topsoil, especially in the latter group, probably originates, at least in part, from coarser sediments from adjacent sandy soils, whilst the clayey subsoil is probably of alluvial or marine origin.

ACKNOWLEDGEMENTS

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THE INFLUENCE OF LITHOLOGY ON SOLUTIONAL PROCESSES, CHILLAGOE KARST, NORTH QUEENSLAND.

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This paper presents a preliminary account of work carried out in the Chillagoe karst early in 1983. The object of the research was to investigate the effects of variation in lithological factors (texture, composition, surface roughness etc) on solutional processes active on outcropping rock. A longer-term aim of the work is to relate form and process in a model of solutional rock sculpture.

The Chillagoe karst:

The area where fieldwork was carried out lies toward the south-eastern end of the Chillagoe karst belt, about 130 km south west of Cairns. Limestones outcrop in a narrow zone which extends NW and then N into the Mitchell-Palmer karst. Tall limestone towers occur throughout this area, and are separated by areas of alluvium, soil, and limestone pediment.

The limestones are probably Silurian in age (De Keyser & Wolff 1964); they have been folded and the bedding now dips almost vertically. Permian granitic intrusions have led to alteration of the limestones in many areas, and marbles of varying crystal size occur in close proximity to relatively unaltered fossil-bearing limestones.

The weathering of these different lithologies has produced a variety of topographic styles. Unrecrystallised limestones are associated with tall, extremely jagged towers; marbles have typically produced lower, more rounded features. The caves and karren of the towers may also be differentiated on the basis of their lithology (Wilson 1976; Dunkerley 1983).

This field situation therefore provides a valuable opportunity to examine the extent to which variations in lithology affect topography and karst solution processes, since climate is essentially uniform through this karst area. Rain is received principally in the months December-March, when about 90% of the annual mean of 800 mm falls in afternoon and evening storms.

Limestone dissolution processes:

The dissolution of a limestone outcrop involves reactions among the three phases: both gaseous carbon dioxide and solid carbonates move into the aqueous phase during this process. There are many factors therefore which can influence the progress of the dissolution reaction, including-

- rate of diffusion of carbon dioxide into the water film
- rate of diffusion of calcium carbonate products into the water film
- water turbulence and flowrate
- temperature and chemical composition of the water
- thickness of the water film
- nature of the rock surface (orientation, curvature, texture, flora)
- rock composition (Ca-Mg-silica-impurities)
- rock texture (crystal size, abundance of lattice defects)

Many of these factors will vary greatly between one body of limestone and another, and many will be influenced by presence of joints, tendency of the rock to undergo granular disintegration , etc. A particularly important influence is exerted by ions of certain metals if these are present in the area of the reaction: the rate of dissolution is retarded, and the apparent equilibrium concentration is reduced (by up to 50% in the presence of Sc ions: Nestaas & Terjesen 1969). Many metals act in this way, including Pb, Sc, Zn, Mn, Mg and Fe. There are several possible explanations for this effect (Nestaas & Terjesen 1969).

Laboratory experiments in which blocks of limestone or limestone powders are dissolved in water can readily only model certain aspects of the field situation.
Further, they are beset by problems such as crystal strain which results from crushing: this may lead to more rapid dissolution than otherwise (Walter & Morse 1984). They can, however, be employed to evaluate the role of inhibitor cations when other conditions are controlled.

In order to obtain data indicative of the relative solubilities of rocks as they behave in the field, or of the rates of dissolution on outcrops of differing lithology, field observations are therefore needed. A proper understanding of solution processes as they operate in the field is clearly important in karst research, and yet there are few field studies of the kind required. The few studies attempting observations on the small scale required to observe solutional processes (eg Jennings 1978) have generally not achieved sufficient resolution in the data collected.

Field methods:

In the present study, two styles of observing site were used:
- simple runoff collectors: These consisted of silastic dams on rock faces with polythene tubing leading to collecting bottles. Dams were installed to collect water samples having travelled various distances across a particular rock outcrop, and on several lithologies.

- Continuous recording sites: At these sites, the water was led from silastic dams via pplythene tubing to flow-through sensor assemblies where rate of flow, temperature, pH, and electrical conductivity were recorded at 1-minute intervals by an automatic data logging system. Three sites were observed simultaneously, the sites being selected to sample water flowing on various gradients and over various distances. Rainfall was also continuously recorded at such sites.

Samples from both styles of site were returned to a field camp where total and calcium hardness, as well as alkalinity, were determined by conventional titration; pH was also accurately observed. Samples were also collected periodically from solution pans and from springs and streams.

Three sites were employed: on fossiliferous limestone at Royal Arch bluff; on moderately coarsely crystalline marble at Racecourse tower, and on coarse marble at Dome Rock.

Saturation indices were calculated from the field data using the WATSPEC aqueous solution model (Wigley 1977).

Results:

More than 40 water samples were analysed during what proved to be an almost rainless wet season. Mean values for selected parameters are listed below-

Site	Mean total hardness (mg/l as $CaCO_3$)	mean pH	mean SIc	slope	grain size
Royal Arch	36.0	8.32	-0.40	64	fine
Racecourse	26.5	8.31	-0.56	56	medium
Dome Rock	22.5	7.93	-0.75	45	coarse

It can be seen that the greatest mean hardness and nearest approach to saturation were achieved on fine-grained rocks at Royal Arch, decreasing hardness and greater undersaturation being associated with progressively coarser rocks. Sharpness of outcrops, and abundance and size of karren fall in the same sequence (Dunkerley 1983).

Results from the continuous recording sites are as yet largely unprocessed. Data which have been examined indicate that for Royal Arch, pH values lie in the range 7.0-9.5; pH rises and calcium carbonate content falls as runoff rate increases. However, neither parameter shows great sensitivity to flow rate, indicating perhaps a mixed surface reaction rate/transport limited control on the export of carbonates. When data from sites where the water has had contrasting residence times on the outcrop have been processed, it will be possible to make firmer statements about this.

Dreybrodt (1981) has evaluated published data on calcite dissolution from laboratory experiments to suggest that solution rates are a function of the ratio of the volume of water to the surface area of calcite. He argues that for large ratios, solution rate is controlled by surface effects and for small ratios, by the conversion of carbon dioxide into carbonic acid. Specifically, if

V = volume of water and
F = surface area of calcite exposed to attack

then where $V/F > 0.04$ there is surface reaction rate control

and if $V/F < 0.01$ dissolution is controlled by hydration of carbon dioxide.

Now these ratios span the range to be expected in dissolution of outcrops by water films, so that indeed mixed rate control could be expected. In runnels of any kind, surface reaction rates will be limiting; on steeper parts of outcrops, with thinner films of water, hydration of carbon dioxide will restrict solution rates.

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RECENT CHANNEL CHANGES AND THEIR RELEVANCE TO RIVER MANAGEMENT: A CASE STUDY OF DAIRY ARM, HUNTER VALLEY, N.S.W.

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INTRODUCTION

Wollombi Brook and some of its tributaries experienced a metamorphosis or complete transformation of channel form during the June 1949 and subsequent floods (for example see Page, 1972). Significant channel widening and bank erosion were recorded. Since then, the N.S.W. Water Resources Commission has implemented a number of river management schemes to stabilise the river channel by reducing the incidence and extent of river bed and bank erosion. Formulation of river management works suitable for this purpose has been hindered by lack of data on the nature, magnitude and causes of recent channel changes. No vertical aerial photography, large scale topographic maps or river gauging information exist for the upper reaches of Wollombi Brook prior to 1950.

The Water Resources Commission (1981) has recently commenced river management works on Dairy Arm, a tributary of Wollombi Brook. We have attempted to reconstruct the channel changes in the Dairy Arm catchment since European settlement using oral and written histories, field surveys and measurements, and stratigraphic descriptions and radiocarbon dating of valley-fill and alluvial fan sediments. The purpose of this study is to determine recent channel changes and assess the suitability of the proposed river management works for river and floodplain stabilisation.

STUDY AREA

Dairy Arm is a right bank tributary of Wollombi Brook in the extreme south of the Hunter Valley and has a catchment area of 39.8 km^2. It drains a catchment of intercalated sandstones and shales of Triassic age (Hanlon *et at.*, 1953). Numerous mass movements are found on or immediately below a prominent structural bench near the confluence of Dairy and Olney Arms (Fig.1). Catchment relief ranges between about 70 and 300m. Mean annual rainfall is approximately 950mm and exhibits a marked summer maximum. Although most of the basin was cleared of forest last century secondary forest has now regenerated except on the valley floor and footslopes in the lower and middle reaches of the valley.

RIVER MANAGEMENT WORKS

The area treated covers the lower 5.5 km of Dairy Arm from about cross section 8 to the Wollombi Brook junction (Fig.1). According to Water Resources Commission (1981) the primary aims of the work were to: i) stop bank erosion; ii) reduce rates of overbank deposition; and iii) provide a stable channel. A controlled alignment conforming to the natural channel pattern and of approximately the same width as the natural channel at bankfull was installed by the following works. Grids of wire mesh fences with willow (*Salix* spp.) plantings (see Rankin, 1982) were installed on the outside of eroding bends and stockproof fencing was constructed elsewhere along large sections of natural bank to allow the regeneration of riparian vegetation. All trees and large organic debris within the alignment were removed. The budget for the works is $183,000. Further details are contained in Water Resources Commission (1981).

CHANNEL AND VALLEY FORM

Fig.2 shows downstream changes in the channel and valley morphology of Dairy Arm. Cross sections 1 and 2 are representative of the upper alluvial reaches where the valley is confined by a structural bench (one of the sandstone members of the Gosford Formation of Hanlon et al., 1953). The channel has a small capacity with a low genetic floodplain covered by a moist forest with Eucalyptus saligna emergents. Immediately downstream of cross section 2 the valley starts to widen and the channel becomes much larger (cross sections 3 and 4 in Fig.2). There are a series of secondary knickpoints (Fig.1) and log steps in this reach resulting in a stepped longitudinal profile. Cross section 4 depicts the channel-in-channel form produced by small upstream migrating secondary knickpoints. Riparian vegetation has been extensively cleared downstream of cross section 3. Cross sections 5, 6 and 7 show a small channel and low floodplain developed within a large trench, predominantly cut in valley-fill sediments. The depth of this trench progressively decreases below cross section 7 until the terrace loses its surface expression immediately downstream of cross section 9. Below the confluence with Olney Arm the valley widens substantially (compare cross sections 10 and 11) and there is an extensive, low, genetic floodplain (cross sections 11 and 13).

RECENT CHANNEL CHANGES

Channel changes have been determined from interviews with long-term residents of the area, portion plans, parish maps, historical documents held by the State Archives and field evidence. Early this century the channel of lower Dairy Arm near cross section 13 was narrow and shallow with occasional large pools lined by reeds. At the Olney Arm confluence there was no channel and wells had to be dug in the valley floor for stock water. Some of these wells still exist. A small sand-bed channel, however was present on the upper reaches of Dairy and Olney Arms.

The June 1949 flood initiated large scale trenching and widening of Dairy Arm upstream of Cullys Arm and Olney Arm upstream of Dairy Arm (Fig.1). Since 1949, landholders recall up to 6m of incision by the upstream migration of a system of knickpoints. On Dairy Arm about 5 km of channel was trenched between cross sections 3 and 8. Knickpoint rotation, however, has occurred near cross section 3 for the following reasons. Decreasing catchment area results in a decrease in discharge and hence both competence and channel size. Larger organic debris (limbs and trunks) can then span the channel in this reach without getting outflanked or dislodged by floods. Some Melaleuca spp. and Eucalyptus spp. have also coppiced and regrown after being felled. The organic debris has formed log steps that act as natural drop structures. On Olney Arm about 5 km of channel has also been trenched. The upstream limit of incision corresponds to a zone of boulder clogs where the channel flows over a structural bench. Immediately upstream of the gullied section is an ingrown meander cave (Jennings, 1970) cut into intercalated sandstones and shales of the Gosford Formation. The maximum height of the cave is 2.5m and its maximum width is 5m, clearly indicating that the channel has been at its present location and elevation for a considerable period of time. A boulder clog also marks the upstream limit of the gully on the unnamed tributary 1 (Fig.1).

Channel incision on Dairy and Olney Arms rejuvenated tributary channels. Knickpoints are still present on Cullys and McMullens Arms

and the unnamed tributary 2 (Fig.1). Fig.3 shows the cross-sectional valley form and stratigraphy of the lower reaches of the unnamed tributary 2. There have been at least two episodes of trenching here, the first phase cut through a high level alluvial fan and the second and most recent, cut through the inset fill deposited within the initial trench. The first phase of incision occurred before 1100 ± 70 B.P. (Beta - 7415), the age of the basal gravels of the inset fill (Fig.3). Prior to the second phase of incision in 1949 the surface of the inset fill was cleared of all trees and the stumps burnt (there are 15 stumps in the lower 100m of this gully). This surface was then buried by up to 0.3m of stratified sands. We believe that the tree clearance and accelerated sedimentation were due to human activities in the catchment after European settlement.

The massive volumes of sediment removed from the upper reaches of Dairy and Olney Arms and their tributaries were largely deposited as channel-fill and overbank sands on the lower reaches of Dairy Arm, downstream of cross section 8. Fig.3 shows the floodplain stratigraphy for cross section 13. Locally there are up to 1m of sandy overbank deposits overlying the pre-1949, finer-grained floodplain soil which was scoured and removed in some places prior to burial (Fig.3). At least one previous phase of valley-wide overbank deposition is also evident (Fig.3) suggesting a correlation with a previous cycle of trenching.

The increased supply of sand from trenching overloaded the channel with bed load resulting in avulsions and cutoffs. The present channel downstream of cross section 13 follows the former road with the old channel abondoned on the opposite side of the floodplain. At cross section 12 a neck cutoff occurred during the March 1978 flood. The abondoned channel is much narrower than the present one. Since March 1978 the channel at cross sections 12 and 13 has degraded up to 0.3m into the bed of the enlarged flood channel.

IMPLICATIONS FOR RIVER MANAGEMENT

The predominant source of sand deposited in the lower reaches of Dairy Arm since 1949 is from valley-fills and alluvial fans that were trenched by upstream migrating knickpoints. Although there are at least thirty mass movements in the Dairy Arm catchment only three of these have supplied any sediment directly to the channel (see Fig.1). These sediment inputs have been small and these mass movements are not an important sediment source.

Significant storage of sand has occurred on the lower floodplain of Dairy Arm and this is clearly not a contributing area of bedload sands. The river management works thus far performed are therefore treating these temporary storage areas and not the primary source. Bank protection works and drop structures located in the upper reaches of Dairy Arm and its tributaries would have been more appropriate. Log steps are important natural drop structures that prevent the upstream extension of a system of knickpoints through small channels. Therefore stream clearing is not warranted, at least in the upper reaches of Dairy Arm.

Further study and field work is required to establish the mechanism of knickpoint initiation in this catchment. Nevertheless it is apparent that any hypothesis must take cognizance of cycles of pre-European trenching.

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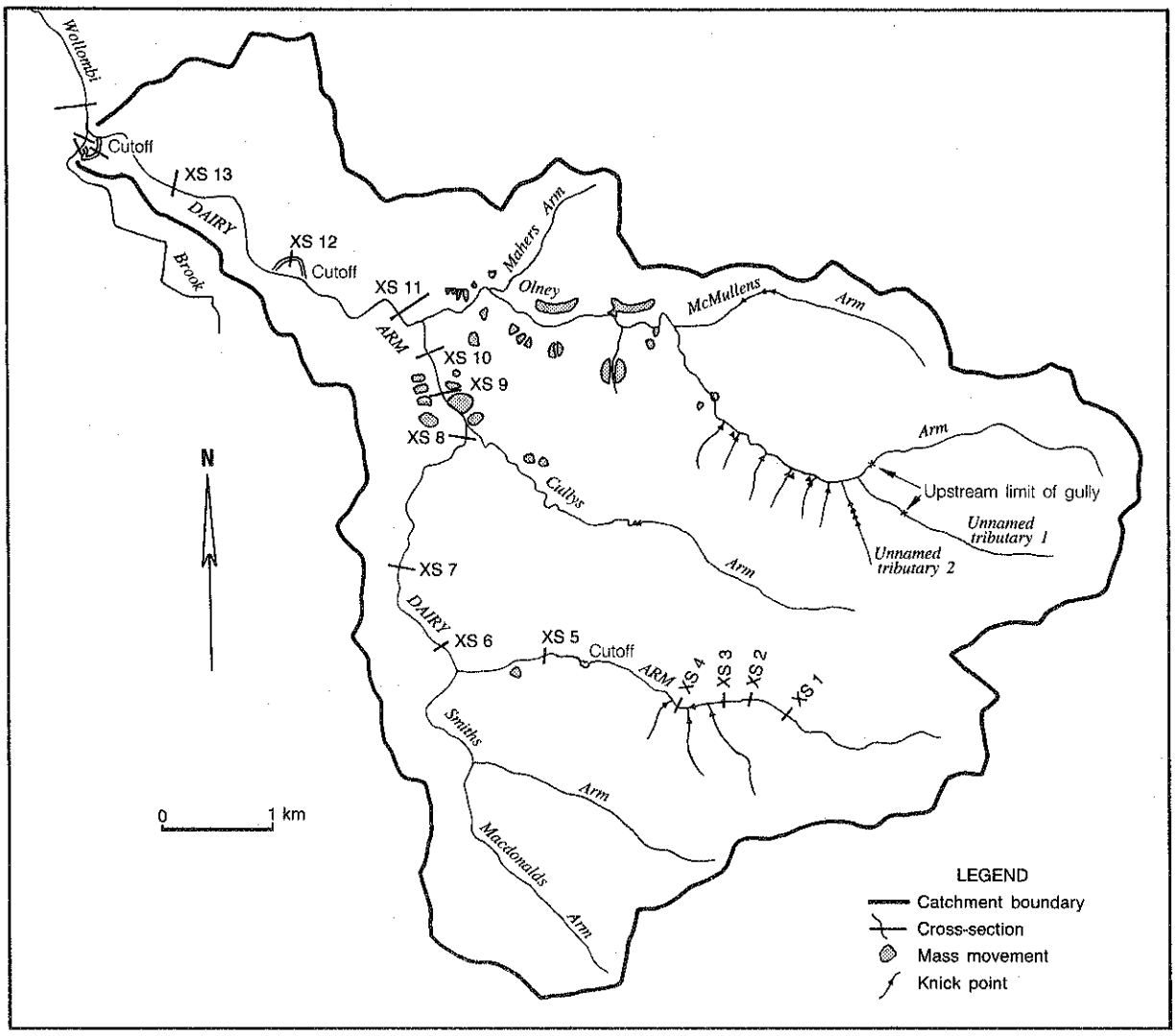

Fig. 1. Catchment of Dairy Arm.

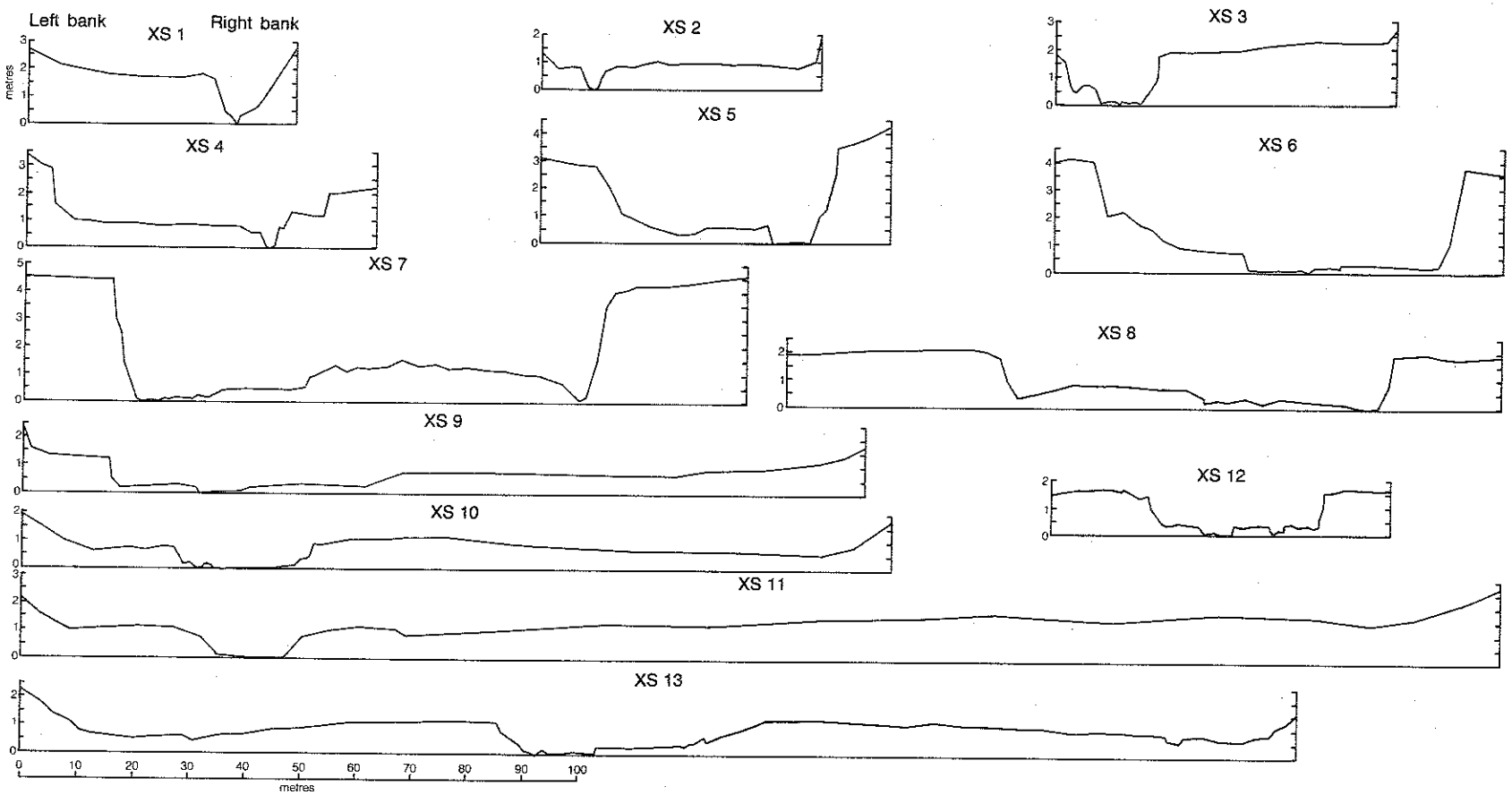

Fig. 2. Channel and valley cross-sections of Dairy Arm.

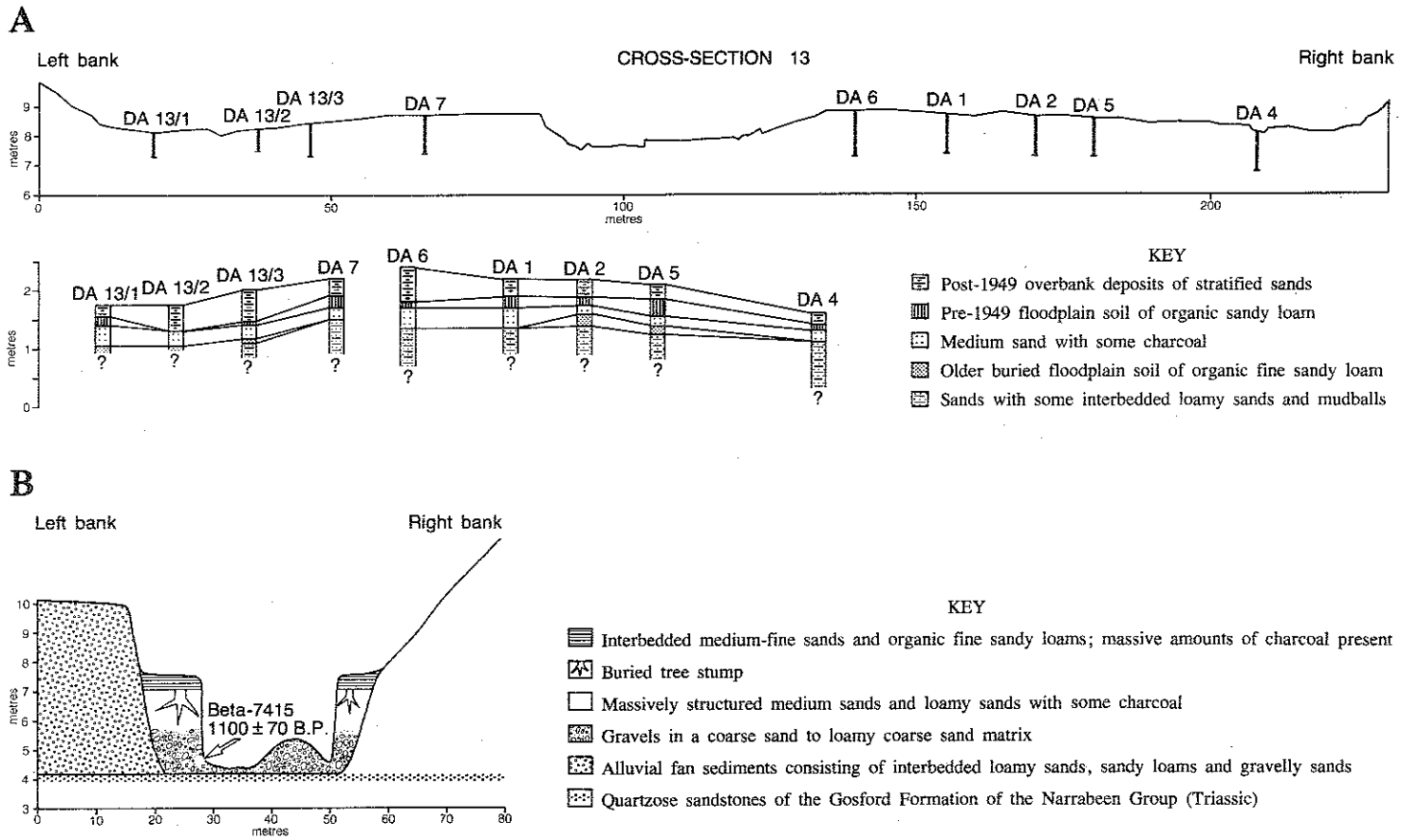

Fig. 3. A: Floodplain stratigraphy of lower Dairy Arm (XS 13).
B: Valley-fill stratigraphy of a tributary of upper Olney Arm.

FOSSIL ICE WEDGES IN PATAGONIA AND THEIR PALAEOCLIMATIC SIGNIFICANCE

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ABSTRACT

The occurrence of fossil ice wedges was studied in gravel pits along a north-south traverse in eastern Patagonia between latitudes 41 and 52 1/2 degrees South. Small wedges were found only south of latitude 50. Their presence implies a rise in mean temperature of 10-12 degrees C since they were formed.

INTRODUCTION

The extent of periglacial features formed during glacial stages has long been of interest as a guide to palaeoclimates. One of the significant aspects is the equatorward limit of former ice wedges (e.g. Poser 1948) which is taken to be an approximate guide to the past distribution of permanently frozen ground. In the southern hemisphere there is little land in the appropriate latitudes and even less with suitable unconsolidated sediments where wedges might have developed during glacial intervals. To my knowledge, no definite ice wedge casts have been reported from Australia and only two rather doubtful instances have been recorded in New Zealand (Te Punga 1957, Washburn 1979). Most of Antarctica is ice-covered and so not available for wedge development but Pewe (1959) has reported sand-filled wedge structures in the McMurdo Sound region.

PREVIOUS WORK

Clearly southern South America is the most likely place in this hemisphere for finding fossil ice wedges and indeed possible ice wedge casts have been reported at a few places. Czajka (1955, p. 139) has briefly mentioned "fossil frostcrack nets" in gravel at 720 m at latitude 46 degrees S in eastern Patagonia but his description is inadequate to judge the validity of this suggestion. Corte (1967) has described small but convincing examples in southern Patagonia. Auer (1970) referred briefly to wedge and pocket structures in central Patagonia but field inspection in 1981 indicated these features may be the result of solution of soil carbonate. Gonzalez and Corte (1976) tabulate a few cases where polygonal surface patterns have been observed but their origin is uncertain. They also describe a system of expanded cracks with up-thrusting of the adjacent flat-lying sandstone and siltstone and tentatively ascribed these features to ice wedge development. However, infilling of a rectangular joint pattern by abundant soil

carbonate seems a more probable explantion of these features. Furthermore, a periglacial origin for these structures would require mean temperatures fully 20 degrees C below those prevailing in this locality today. Liss (1969) described possible fossil ice wedges in gravels at Puerto Madryn in latitude 42 1/2 degrees S but he himself was uncertain of their origin. In 1981 the lower and more critical part of the section was obscured; disturbances in the upper part of the exposure were visibly due to growth and/or dissolution of carbonate.

Thus all reports of ice wedge casts in Patagonia are uncertain apart from Corte's 1967 study. Part of a field trip to Argentina in 1981 was devoted to investigating this question and the results are presented here.

THE PATAGONIAN ENVIRONMENT

Patagonia is a general term for the southern "cone" of South America extending polewards from about latitude 40 degrees S to latitude 56 at Cape Horn. In Australia these latitudes correspond to Bass Strait and c. 150 km south of Macquarie Island. The Andes Mountains fringe the Argentinian portion of Patagonia on the west but most of the area consists of low plateaux incised by wide, flat-floored valleys. Some of the plateaux consist of extensive Tertiary basalt sheets generally overlying weak Cainozoic and Mesozoic sediments but level surfaces underlain by Cainozoic gravel are even more ubiquitous; similar gravel floors the wide valleys. Much of the gravel is outwash from Cainozoic glaciers (Mercer 1976) but in the north at least they must have some other origin, presumably as flood deposits from the mountains to the west. Gravel is a suitable material for the development of ice wedges. Patagonia also has hilly areas on Mesozoic volcanics and basement rocks.

The climate is cool, dry and very windy. Mean annual precipitation is 150-200 mm and fairly evenly distributed throughout the year. January mean temperatures range from the low 20s in the north to the low teens in the south; July means range from about 7 degrees in the north to about 1 in the south. A dry climate may have hindered the growth of ice wedges but they exist today in comparably dry areas of the Arctic.

Glaciers still exist in the higher parts of the Andes. During Pleistocene glacial phases large valley glaciers extended to the foot of the mountains in the north while in the south they increasingly encroached on the lowlands. In the extreme south of the mainland at Rio Gallegos (lat. 52 S) and on Tierra del Fuego, the ice extended at least as far east as the present coast (Fig. 1).

Thus, along a north-south traverse of 1600 km following the main highway in eastern Patagonia, precipitation and altitude are fairly uniform, homogenous gravels are the dominant surface lithology and temperature falls regularly with increasing latitude. These circumstances offer favourable conditions for determining the equatorward limit of Pleistocene fossil ice wedges.

OBSERVATIONS

Along the highway borrow pits conveniently expose the gravels every few tens of km. Some 30 such pits were examined between latitudes 41 and 52 1/2 degrees S.

On the traverse no evidence of cryogenic structures was observed north of latitude 50 degrees. From this point southwards to the Chilean border at latitude 52 1/2 deg. there was increasing evidence for cryoturbation. The traverse did not extend further south to Tierra del Fuego.

The first sign of cryoturbation in the gravel pits was an increase in the proportion of erected pebbles in the top metre (Caillleux and Taylor 1954). Between Santa Cruz and Rio Gallegos, around latitude 51 deg. S, shallow wedges were observed for the first time (Fig. 2) though not in all exposures. They were 70-100 cm deep, 10-20 cm wide, had erected pebbles on their margins and a filling of sand and silt. They were very similar to the features descibed near Rio Gallegos by Corte (1967). At some sites they occurred singly and elsewhere in groups which presumably formed part of a network of tundra polygons.

South of Rio Gallegos (lat. 52), Pleistocene glacial deposits extend to the present coast and consequently the material exposed to possible wedge-forming conditions before and after glacial maxima was likely to be till rather than water-laid gravel and so may have been less suited to wedge development. Nevertheless, the clearest evidence of fossil ice wedges encountered on the entire traverse was found in a roadside pit about 40 km south of Rio Gallegos. Here two generations of weathered till with much carbonate were exposed (Fig. 3). The upper till was overlain by an irregular stone line with many ventifacts which was in turn overlain by 20-40 cm of silty material with a weakly developed soil profile. Wedge casts up to 1 m deep occurred in both tills, with some of those in the upper till extending down into the lower till where the carbonate-enriched material showed signs of up-arching. The upper portions of thelower wedges were distorted in the downslope direction suggesting mass movement of the till after their formation.

It is interesting to note that these relics of Pleistocene periglacial patterned ground occur adjacent to modern patterned ground in the form of linear gilgai on young volcanic soils.

DISCUSSION

The overall picture from mainland Patagonia makes good sense with erected pebbles and small wedge casts appearing at the colder end of the traverse. The small size of the features encourages the belief that they mark the equatorward limit of fossil ice wedges and hence of permafrost in this region while the association with glacial till supports, though does not prove, a cold-climate origin. At the same time the evidence reinforces the doubts expressed above concerning the periglacial nature of structures described by other workers further

north.

Ice wedges are believed to require mean annual temperatures no higher than -5 deg. C (Washburn 1979) although recently examples have been described from Alaska which formed under slightly higher temperatures (-3.5 deg C.; Hamilton et al. 1982). The present mean annual temperature at Rio Gallegos is 7 deg. C and hence the ice wedge casts imply temperatures 10-12 degrees C lower than they are today. Differences from modern temperatures were not necessarily the same at all seasons. In the absence of dating it is sensible to ascribe the ice wedge casts to the maximum of the last glaciation around 18-20 000 years ago. The lower wedges illustrated in Figure 3 presumably date from an earlier glacial maximum.

The implied temperature shift of around 11 degrees C since full-glacial times is close to that postulated for the Snowy Mountians of New South Wales and the south-western United States (Galloway 1965, 1970, 1983a). These other estimates were based on reconstructing the timber line position from what were believed to be periglacial solifluction deposits.

Mean annual temperatures 11 degrees lower than they are today would not suffice for ice wedges to develop in Tasmania and would scarcely permit their formation in the South Island of New Zealand which is in accord with the available field evidence. On the other hand, such lower temperatures should have sufficed for ice wedges to develop in Tierra del Fuego and the Falkland Islands provided suitable lithologies occurred and potential sites were not subsequently drowned by the Flandrian transression. These localities should therefore offer a good opportunity for checking the ideas expressed here.

There is a persistent belief that periglacial phenomena are caused by the presence of glaciers. (e.g. Wright 1982). In fact the reverse is more nearly true: glaciers form under a sub-set of periglacial climates. There is no reason to believe that the former presence of glaciers as far east as Rio Gallegos promoted the development of ice wedges in this locality. Indeed, the climatic situation is such that relatively warm foehn winds would have occurred in the lee of the ice.

The presence of fossil ice wedges in southern Patagonia provides no information on the ice age precipitation other than demonstrating that the area was not totally without water. Pollen studies in Chile in the latitude of northern Patagonia (Heusser et al. 1981) suggest that full-glacial conditions were drier than now and mean temperatures about 5 degrees C lower. While conditions west of the Andes may not be entirely relevant to Argentinian Patagonia, this evidence does suggest that the ice wedges developed under a dry cold climate. If this is so, the vast former lakes in Patagonia (Galloway 1983b) probably were not full-glacial in age because even the drastic reduction in evaporation associated with temperatures 10-12 degrees lower would hardly have sufficed to maintain them unless there were also a concomitant increase in precipitation. Markgraf (1983) reports pollen evidence for cool, wet conditions in eastern Tierra del Fuego around 10 000 yr B.P.

The evidence presented here indicates that glacial/modern temperature contrasts were about the same in both hemispheres. A similar conclusion was reached from study of the glaciation in the Snowy Mountains of New South Wales (Galloway 1963; Galloway et al. 1973, p. 132). The vast difference in the extent of ice in the two hemispheres then and now did not differentially affect the Pleistocene temperature changes.

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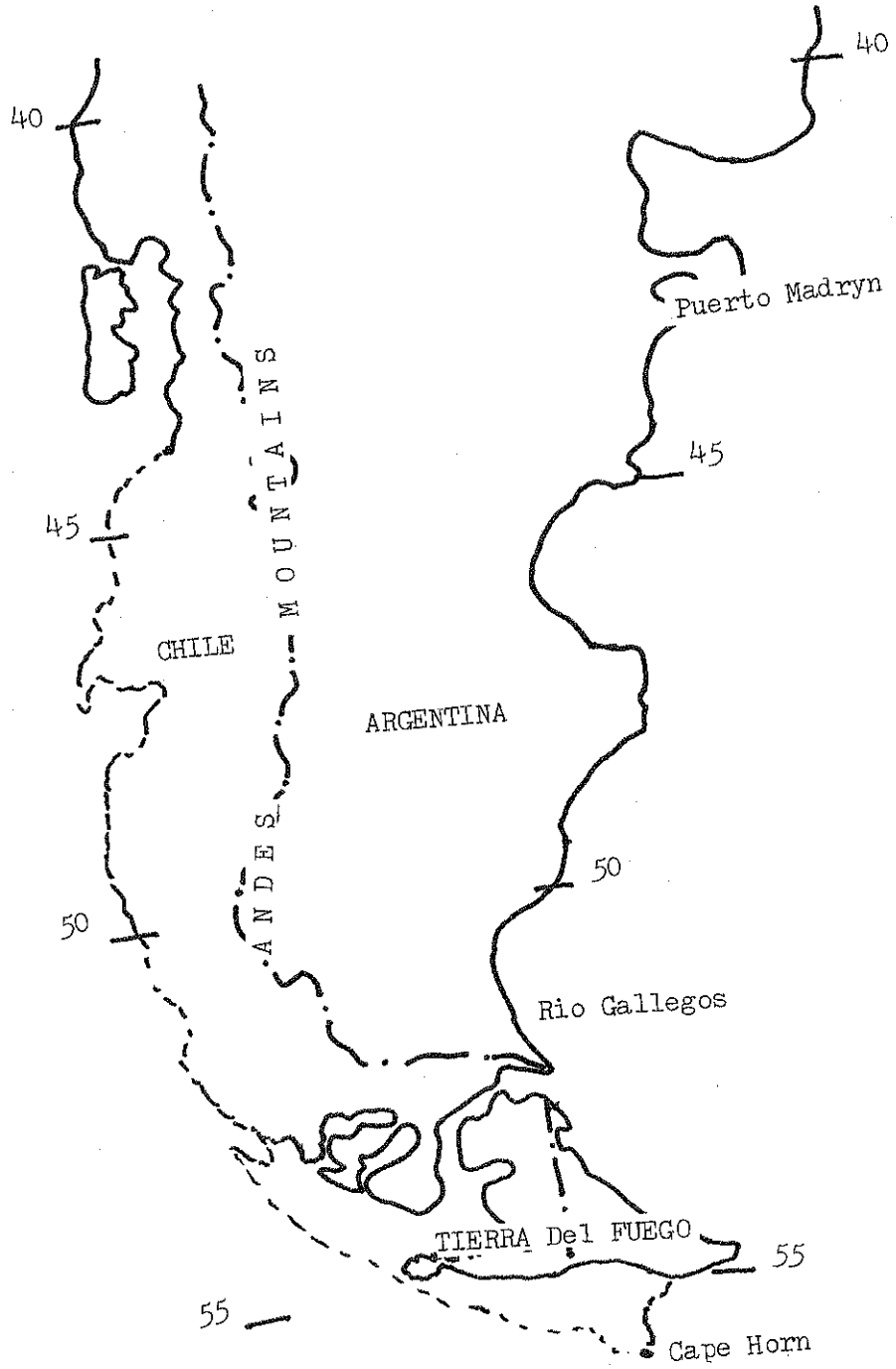
40
40
Puerto Madryn
45
45
CHILE
ANDES MOUNTAINS
ARGENTINA
50
50
Rio Gallegos
TIERRA Del FUEGO
55
55
Cape Horn

FIGURE 1.

WEDGE CAST IN PATAGONIAN GRAVEL c. Lat. 51°S.

1. Soil
2. ventifacts
3. wedge with silt/sand fill
4. patches of soil carbonate

FIGURE 2.

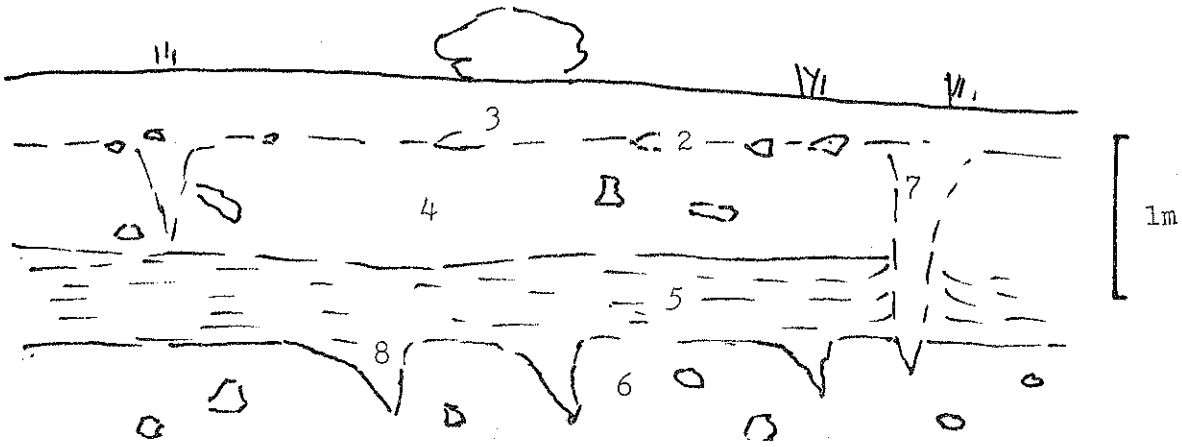

FIGURE 3. WEDGE CASTS IN TILL c. Lat. 52°30'S.

1. soil
2. stone line
3. ventifact
4. upper till
5. heavy carbonate layer
6. lower till
7. younger wedges
8. older wedges

PROBLEMS IN THE APPLICATION OF BEDLOAD FORMULAE TO N.S.W. COASTAL RIVERS

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ABSTRACT

Eight bedload formulae suitable for gravel bed streams are applied to ten rivers located in four large coastal drainage basins of N.S.W. Practical and theoretical problems of implementation of the formulae are discussed as well as the variability of the results obtained. It is concluded that studies involving estimates of bedload on N.S.W. coastal rivers should preferably avoid the use of bedload formulae because of inadequate river gauging data, the lack of consistancy of the formulae and the limited supply of sediment to these rivers.

INTRODUCTION

Bedload formulae have found considerable use amongst geomorphologists and engineers for they provide a rapid and cheap means of estimating bedload yields. Typical applications are in connection with dam sedimentation, gravel and sand extraction and studies of channel degradation.

Because it is widely acknowledged that certain bedload formulae may be highly inaccurate (eg White, Milli and Crabb, 1975; Church, 1975; Vanoni 1975; Hudson, 1983, they are usually selected on the basis of two important criteria:
1) Whether the formulae were developed under a range of conditions (slope, grain size etc) similar to those in the river(s) to be evaluated.
2) Evidence of a successfull application of the formula in a natural river where actual bedload yields have been measured directly.

The second criteria is impossible to satisfy for N.S.W. coastal rivers owing to the lack of actual bedload measurements in these streams. This study looks briefly at the performance of different formulae chosen only on the basis of the first criterion. Those that give results that are inconsistant with geomorphological evidence are identified as well as those that give highly variable results between sites when there is no evidence for such variation. Other problems associated with the application of the formulae to N.S.W. coastal rivers are also discussed.

THE STUDY AREAS

Bedload calculations were made for a total of 11 stations in 4 coastal river basins of N.S.W. gauged by the Water Resources Commission. These river basins were the Manning, Kuruah, Shoalhaven and Clyde. Table 1 shows the characteristics of each site.

The physiography of the drainage basins varies appreciably but all contain substantial areas of deeply disected topography in parts of their catchments. The sources are for the main part just inland from the eastern edge of the highlands at altitudes ranging from 600-1500m. Main streams have only their headwaters in the highlands except in the case of the Shoalhaven where about 1/3 of the basin is situated in the highlands.

Rainfall for the Manning , Karuah, and Clyde basins averages around 1000mm while that for the Shoalhaven is substantially less than this. The river basins are comprised mainly of Paleozoic fold belt strata with some more recent sedimentary facies and volcanics present in certain locations. Forest covers approximately 65% of the Manning, 60% of the Karuah, 55% of the Shoalhaven and 95% of the Clyde basin, with the remainder of each catchment in pasture. The drainage basins in all cases are characterised by paved gravel bed streams and thus only formulae suitable for gravel bed conditions can be used.

BEDLOAD FORMULAE

The bedload formulae selected for the study were the Einstein Bedload Function (Einstein 1950), Meyer-Peter and Muller (1948), Schoklitsch (Shulits 1935), Engelund and Hansen (1967), Modified Einstein (Colby and Hubbell 1961), Einstein-Brown (Brown 1950), Bagnold (1980) and Parker (1982) equations. Reviews of most of these formulae may be found in Vanoni (1975) and Simons and Senturk (1977). All the formulae have either been suggested, reviewed or used by other workers for gravel bed conditions (eg: Henderson, 1960; Church, 1975; Hudson, 1983).

METHODS

Details of field methods are described in Hean (1983). All hydraulic geometry data for the flows at each site were made available by the Water Resources Commission of N.S.W. Annual bedload yields were computed from the values given by each bedload equation according to the method of Piest (1964). These calculated estimates for each equation were then compared with each other. geomorphological evidence (on the availability and likely flux of coarse sediment in each river) was gathered from examination of aerial photographs and qualitative field observations.

RESULTS AND DISCUSSION:

CONSISTANCY OF FORMULAE AND PROBLEMS OF IMPLEMENTATION

Table 2 shows annual bedload yields for each of the 11 sites and 8 equations. It is immediately apparent that there are staggering variations in the results from using different equations at the same site, particularly for the Killawarra and Welcome Reef sites which are both on trunk streams of large drainage basins. The results between sites for the same equation also show marked inconsistancies. The Meyer Peter and Muller equation is a good example. This equation which is generally considered to be one of the more reliable for gravel bed rivers and was reported by Pickup and Higgins (1979) to give excellent results for a river transporting an abundant gravel load in Papua New Guinea. However, it gives zero yields for six of the sites in this study and yet appreciable yields on another three sites. In all, these results imply that no single formulae can be expected to perform adequately under a wide range of conditions, a conclusion also drawn by Hudson (1983).

However, the poor performance of individual formulae may also be due more to the poor quality hydraulic geometry data available for the gauging sites on these rivers than to inherent errors in the formulae. In most cases the hydraulic geometry measurements were for flows less than 10% of the highest flow indicated by flow duration curve data. This

necessitated the extrapolation of velocity and mean depth measurements so that bedload could be calculated over the full range of flows to produce a bedload rating curve. Although power function relationships are commonly used in assessing at-a-station changes in hydraulic geometry (Leopold, Wolman and Miller (1964), this may be a significant source of error in the bedload calculations. A simple check on the accuracy of such extrapolations was done using the Killawarra site, a particularly well gauged site where hydraulic geometry has been measured for a flow 40% of the highest ever recorded. Extrapolation of data collected for flows up to 10% of the maximum flow ever recorded resulted in a 20% error for hydraulic geometry parameters at flows of 40% of the highest. A further error resulting from Water Resources Commission data is that channel geometry measurements and associated flow data are not collected from exactly the same cross section for successive flow events at each gauging site. Depending on the flow conditions, guagings are taken at a number of possible cross sections within a gauging reach. When these hydraulic geometry measurments at various points were incorporated in the above analysis the extrapolated results were in error up to 65%. Because extrapolation of hydraulic geometry data measured at more than one point is the rule rather than the exception in N.S.W. rivers, these errors are a major problem with the application of bedload formulae to these coastal rivers

WITHIN BASIN CONSISTANCY OF FORMULAE

The sites in the Manning basin provide an opportunity to comment on within basin consistancy of the bedload formulae. Above the Killawarra site, the Manning river is fed by four major streams for which bedload yields were calculated at the Doon Ayre, Woko, Mackay and Rocks Crossing sites. The combined drainage areas of these sites is 86.5% of that at Killawarra. Similarly combined annual discharges are 83.7% of that at Killawarra. Sediment continuity would imply that the summation of the bedload yields on the tributaries should be of the order of 85% of the bedload yield on the trunk stream. However, it is clear from Table 3 that none of the formulae come anywhere near accomplishing this proportion. They all completely violate the conservation of sediment volume. Depending on the choice of formulae, massive channel aggradation or degradation is implied, yet there was no geomorphological evidence for either.

CALCULATED BEDLOAD TRANSPORT RATES COMPARED WITH GEOMORPHOLOGICAL EVIDENCE

Of the four drainage basins in the study the Manning was examined in the most detail. Here field reconnaisance and aerial photographs indicate that the rate of contribution of bedload-sized material to the river is very slow. There appear to be no obvious large sediment sources. Although steep slopes prevail over much of the basin they are almost without exception well vegetated and stable. There is little evidence of any rapid lateral movements of channels in the Manning basin indicating that even the reworking of alluvial deposits is of limited significance. Major portions of the rivers in the Manning basin are confined to narrow valleys where large volumes of floodplain and terrace material are not present.

Table 1 CHARACTERISTICS OF STUDY SITES

River Basin	River	Site	Drainage Area	% of Basin	Annual Discharge	Flow Record Length
Manning	Manning	Killawarra	6560 sq. km	78.1	2 035 800 ML	1945-82 37 years
	Nowendoc	Rocks Crossing	1840	21.9	626 055	1946-81 35
	Gloucester	Doone Ayre	1590	18.9	617 242	1945-81 36
	Little Manning	Woko	470	5.6	113 600	1964-81 17
	Barnard	Mackay #2	1770	21.1	345 000	1963-81 19
Karuah	Karuah	Booral	974	44.7	358 700	1968-81 13
	Karuah	Monkerai	203	9.3	146 500	1945-81 36
Shoalhaven	Shoalhaven	Welcome Reef	2770	37.9	802 500	1929-79 51
	Mongarlowe	Marlowe	417	5.7	172 000	1945-80 35
	Crang	Hockeys	156	2.3	81 000	1924-79 55
Clyde	Clyde	Brooman	860	47.8	553 100	1960-82 22

Table 2

MEAN ANNUAL YIELDS OF BEDLOAD FOR SITES AND EQUATIONS (TONNES)

Site	MPM	Eng.	Bag.	Park.	Schok.	Mod. Ein	Ein.	Ein-Brown
Kill.	12,244	110,077	13,189	792,890	27,590	235	1,179,698	1,588,608
Doon	-	345	152	-	309	-	1,132	725
Woko	-	425	-	-	108	-	1,643	791
Mack.	3,604	5,346	1,035	930	1,756	402	7,600	7,776
Rock.	59,904	24,082	2,429	37	1,842	18,168	17,635	13,733
Monk.	-	230	419	115	743	-	1,742	4,032
Boor.	-	-	-	-	-	-	-	-
Broo.	-	3,719	246	1,355	863	-	17,269	15,165
Mack.	-	5,095	1,816	31,519	5,127	-	83,707	59,512
Welc.	22,971	298,555	60,438	3,052,852	215,743	119	4,006,369	14,914,101
Hocks.	513	1,151	1,600	273	1,808	33	1,293	1,841

Table 3 CONSISTENCY OF FORMULAE WITHIN A BASIN

SUM OF TRIBUTARY SEDIMENT YIELDS TO THE MANNING AT KILLAWARRA (TONNES)

	MPM	Eng	Bag	Park	Schok.	Mod.Ein.	Ein.	Ein.Brown
Above Killawarra	63,508	30,198	3,616	967	4015	18,570	28,010	23,025
Killawarra	12,244	110,077	13,189	792,890	27,590	235	1,179,698	1,588,608
% of Value at Killawarra	519%	27%	27%	0.12%	14.6%	7902%	2.37%	1.45%
Value 85% Killawarra	10,407	93,566	11,211	673,957	23,452	200	1,002,743	1,350,317
Assuming combined totals should be 85% Killawarra								
Factor by which each equation is out	6.10	3.1	3.1	697	5.84	93	36	59

Gravel bars along the rivers are shallow and bedrock frequently protrudes to the surface. Vegetation colonizes many of these bars and substantial armouring is present particularly on the more exposed bars. Typical size ratios of surface to subsurface material range from 1.5-3.0 for the areas sampled. This indicates that even within channel sources of bedload are quite restricted. Mean surficial bedload size in the Barrington-Gloucester system (the Manning's largest tributary) decreases rapidly over the first 50km to only 1/4 of that in the headwaters. Similarly, Jenks (1982) has shown that there is a rapid dimunition in mean sediment size from near the Killawarra gauge downstream for 35km to Dumeresq Island in the Manning estuary. Such reductions in sediment size through attrition under conditions of abundant bedload transport is most unlikely over comparitively short distances. Similar arguments can be presented for the Karuah, Shoalhaven and Clyde river basins which show even greater armouring. This evidence of very limited coarse-sediment load is also consistent with the growing body of recent evidence showing that rates of denudation on the eastern highlands (eg Young, 1983) and the coastal plain (eg Young and McDougall, 1982) are extremely slow. Furthermore, these landscapes are exceedingly old and stable (eg Wellman and McDougall, 1974; Young, 1977; 1981; Francis and Walker, 1978; Bishop Hunt and Schmidt, 1982).

The dependence of bedload yield on sediment supply has been amply demonstrated by Nanson (1974). Clearly, if geomorphological evidence for restricted coarse-sediment supply is correct then there is little basis for the application of bedload formulae to these rivers. The most crucial assumption of bedload formulae is that there is unlimited sediment availability and that the whole channel bed is potentially mobile. This assumption is obviously violated for N.S.W. coastal rivers as large segments of the streams are cut in bedrock and have only partially alluvial and often well armoured boundaries.

The bedload formulae have been developed for mid to high latitude northern hemisphere streams where glacial-perglacial conditions have left a legacy of abundant coarse sediment still being actively eroded from river basins. These conditions do not apply to east coast N.S.W. rivers which are probably operating far below the calculatable maximum bedload yield because of limited sediment supply. However, those formulae (Table 2) giving small results that are in best agreement with geomorphological evidence of limited sediment supply may in fact be the least accurate in hydraulic terms because they were developed for conditions of unlimited sediment supply.

CONCLUSION

The present study has identified several major problems with the application of bedload formulae to N.S.W. rivers.

1. The absence of hydraulic data for anything like a full range of flows at gauging stations necessitates extrapolation of channel geometry data from very low flow measurments.
2. The lack of consistancy between results for a range of formulae at a single site and the inability of any of the formulae tested to provide downstream sediment-yield continuity for sites in the same river basin. Both suggest that choosing formulae whose flume studies for the main part used gravel sized sediment and which have been reviewed for

use in gravel rivers is no guarantee of gaining a result that is even of the correct order of magnitude! Only proven comparitive studies of the formulae with measured data will suffice and yet, as previously stated, no such data is available for east coast rivers in Australia.

3. The assumption of unlimited sediment availability is invalid for the rivers in this study. As a consequence, existing formulae are inapplicable. Furthermore, the dominant influence of limited sediment availability makes impossible the development of new hydraulically based formulae for these conditions.

When making estimates of bedload transport for gravel bed N.S.W. coastal streams it would seem prudent to avoid reliance on bedload formulae and to place greater emphasis on detailed geomorphological investigations.

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SAND AVAILABILITY, EQUIVALENT SAND THICKNESS AND LONGITUDINAL DUNE PATTERNS IN THE NORTHWESTERN SIMPSON DESERT, CENTRAL AUSTRALIA

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ABSTRACT

A discussion of relationships between sand availability and 'equivalent sand thickness' (EST) resulted in a statement that these tend to be practically synonymous in fields of longitudinal dunes, where little sand normally underlies the swales. This paper describes the existence in the northwestern Simpson Desert of three contrasting longitudinal dune patterns, claimed to be expressive of differences in the degree of contrast in sand transmissivity between swale and dune surfaces. In one pattern, suggested as having been formed by wind-sculpting of alluvial sands *in situ*, the equilibrium pattern stage of dune development, as determined by sand transmissivity controls, appears to have been reached with a limited depth of wind-sculpting, leaving potentially erodible sands beneath the dune floor. In the other two patterns, in which dune sand has been entirely or in part wind-transported into the areas, EST is equivalent to sand availability within a range of equilibrium dune patterns set by ground-surface controls.

INTRODUCTION

In 1977, on the basis of consideration of the distribution and relationships of three simple dune types (barkhans, longitudinal and transverse dunes) I concluded '... The three main simple dune forms are seen to be differentiated by the amount of available sand and its potential transport, rather than by wind regime, with the progression from barkhan through longitudinal to transverse dunes marking an increase in the thickness of sand and a relative diminution in the extent of surfaces suited for its rapid transmission'. In the case of longitudinal dunes it was postulated ... 'Longitudinal dune systems require a particular combination of sand supply and terrain. The sand should not be so plentiful as to mask the downwind-trending surface contrast between ridge and swale, nor too scanty to allow the growth of continuous forms' (Mabbutt, 1977)

Wasson and Hyde (1983) have considered the relative importance of sand availability and wind regime in determining four elemental dune types (including star dunes as a fourth type). They conclude that 'longitudinal dunes occur where the winds are more (than moderately) variable but there is little sand'. As a quantitative measure of the availability of sand they used 'equivalent sand thickness (EST) defined as the thickness of a continuous sheet of sand which results from the hypothetical spreading out of the dunes, over a specified area'. They noted that EST and sand supply are often but not always identical.

In a discussion of that paper, Rubin (in press) listed five ways in which 'sand availability' might be used:

- the volume of sand within dunes in an area of dunefield (which he equates with EST as proposed by Wasson and Hyde);
- the total value of potentially erodible sand in a dunefield, including sand beneath the swales;
- the fraction of dunefield covered by erodible sand;
- the rate at which sand is supplied from upwind to a unit width of dunefield;
- the net rate of sand accumulation in a dunefield.

In their reply, Wasson and Hyde (in press) note that EST may not account for all the aeolian sand in the dunefield, but state that 'very little (aeolian sand) lies in (swales) in fields of longitudinal dunes'.

This paper briefly considers the question of sand availability and ground-surface character in relation to longitudinal dune patterns in the northwestern Simpson Desert.

LONGITUDINAL DUNE PATTERNS AND THEIR PHYSIOGRAPHIC SETTINGS

A recent analysis of the plan geometry of longitudinal dune patterns in the northwestern Simpson Desert (Mabbutt and Wooding, 1983) claimed to identify three main patterns on the basis of dune form and mean spacing, numbers of terminals and segments, dune segment frequency and mean length, connectivity and avoidance (Fig.1). Type I pattern consists of closely-spaced dunes (200-400m) with sandy swales of catenary cross-section, commonly with significant differences in elevation between adjacent swales. It is characterized by extremely high dune segment frequencies, high connectivity and by low mean segment length values. Type II dunes, with spacing in the range 400-800m have flat-floored swales with a very thin loose sand cover over compact sandy alluvium. Type II patterns show intermediate values for dune segment frequency, connectivity and mean dune segment length. Type III dunes, with spacing greater than 800m, are generally separated by sand-free stony corridors; they lack the contrast between vegetated plinth and mobile sandy crest present in types I and II dunes. This pattern exhibits extremely low dune segment frequencies, low connectivity values and very low mean dune segment length. All three patterns are judged to be expressive of a dynamic equilibrium state locally, in that the numbers of ridges crossing the upwind and downwind limits of sample areas are approximately equal.

Dunes of type I pattern occur adjacent to and immediately downwind from active or prior river channels and their sandy alluvial flood-outs. Type II pattern dunes, which are more widespread, occur mainly downwind from type I at greater distances from areas of evident fluvial deposition, and rest on an alluvial surface, in part gravelly, which extends beneath much of their area of occurrence and slopes up gradually northwards towards the desert margin. Type III pattern dunes mainly occur where type II dunes are interrupted by

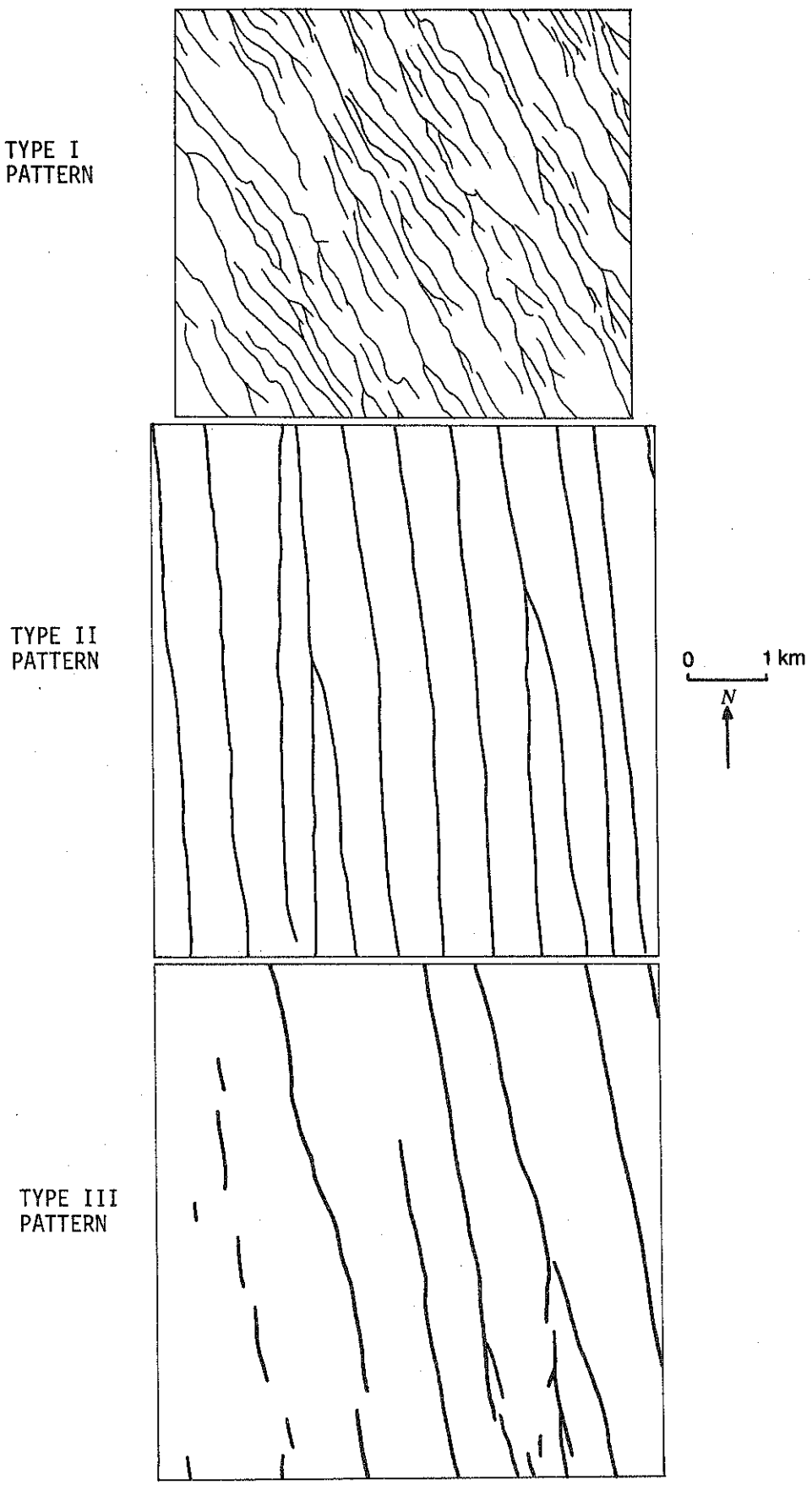

Fig. 1. Longitudinal dune patterns in the northwestern Simpson Desert (from Mabbutt and Wooding, 1983)

broad gibber-strewn rises in the desert floor.

ORIGINS OF THE DUNE PATTERNS

Type I pattern dunes are considered to have originated in areas of prior alluvial deposition in two main ways:

- by wind-fashioning of alluvial sands
- as dune chains streaming from sandbed channels.

Total sand depth = 3m

EST values of type I dunes in the study area are ± 2.5m, indicative of shallow aeolian reworking. This is consistent with a dune pattern which is regarded as less organised than that of type II, for example, on the basis of high connectivity values; however the further development of type I dunes was not restricted by sand availability, in that erodible sands are known to occur beneath the present swales.

Type II pattern dunes occur downwind from type I pattern dunes, at greater distances from identifiable fluvial sand sources. They must be assumed to have originated from sand transported from such sources across an underlying alluvial floor (Mabbutt and Sullivan, 1968) and/or from wind-fashioning of overlying deposits related to an earlier, more extensive fluvial domain. Whether the result of dune extension or of wind-fashioning *in situ*, type II pattern dunes represent a more organised system than type I, in which increased spacing has been achieved through the coalescence or downwind termination of ridges. This is associated with an EST of between 4 and 5m across the range dune spacing. What also distinguishes type II pattern dunes is the near-exposure of an underlying firm alluvial surface in the swales.

Type III dunes can be seen to have transgressed into the stony areas in which they now occur, by extension of type II dunes. EST values are in the range of 0.5-1m.

DUNE PATTERN AND SAND AVAILABILITY

The analysis of dune patterns (Mabbutt and Wooding, 1983) stressed the importance of the ground-surface contrast between dune and swale, through resulting differences in sand transmissivity, as a determinant of differences between the longitudinal dune patterns, a conclusion which has some relevance to the question of EST and sand availability. The argument follows, for example, that further development of the type I dune pattern may have been inhibited by the lack of contrast between the dune and swale surfaces, restricting the depth of wind-sculpting, as represented in an EST value of around 2.5m only, although additional underlying sand was available for erosion. The greater mobilization of sand (4-5m EST) represented in the type II pattern was explained by Mabbutt and Wooding as due to the greater surface contrast between the firm swale and the loose sandy dune ridges. This implies that part of the sand represented in the type II dunes was transported into its present area of occurence, across the underlying firm alluvial floor, since any pre-existing sand mantle could not have been so thick as to prevent exposure of that surface by wind-sculpting.

The study referred to supports the suggestion by Wasson and Hyde (in press), that in fields of longitudinal dunes EST is equivalent to sand availability in cases where dunes have formed wholly from transported sand or where an underlying surface of sufficiently contrasting sand transmissivity has been exposed by wind-sculpting. However the evidence from the type I pattern dunes in the north-western Simpson Desert suggests that these conditions may not always be achieved in fields of longitudinal dunes, to the extent that the necessary development of dune formation may be prevented through limited sand transmissivity differences, related to lack of surface contrast between dune and swale, whereby underlying potentially erodible sands remain unexploited.

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DESERT LOESS DEPOSITION PROCESSES IN WEST AFRICA

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ABSTRACT

The results of a process study of the Harmattan wind and its dust, are used to demonstrate that the 20-50μm particle-size criterion taken as typical of all loess deposits, may not be a realistic criterion for identifying desert loess deposits.

INTRODUCTION

Research into the nature and origin of loess has generated more than the usual amount of debate and in particular the origin of desert loess, as distinct from glacial loess, has been a controversial question. Much of the general loess debate has centred upon finding an acceptable definition of what is loess.

Smalley and co-workers (Smalley 1975, Smalley 1978, Smalley and Vita Finzi 1968, Smalley and Krinsley 1978) have made significant contributions to the general loess debate and appear to have largely resolved the definition of typical loess as:

> "Clastic deposit which consists predominantly of quartz particles in the size range 20-50μm, and which occurs as wind laid sheets.

This question of loess definition has, however, received a greater input from the more numerous glacial loess than desert loess workers, with the result that the definition more adequately described glacial loess than desert loess. The response of glacial loess workers (eg. Smalley and Vita Finzi 1968) has been to question whether, in fact, true loess deposits can be formed in and around deserts.

Perhaps the most important constraint upon loess research in general, is the infrequency with which aeolian loess deposition processes operate at present. In the present paper aeolian process evidence from the Harmattan wind in West Africa is used to examine one feature of the loess definition: the 20-50μm size criterion.

THE HARMATTAN WIND

In terms of quantity of dust and distance transported (Goudie 1978), the Harmattan wind constitutes the largest single aeolian system operating on the earths surface at present, and as such represents a valuable case-study of present day aeolian (loess) processes. The principal source of dust appears to be the Bodele Depression of the Chad Basin (Hamilton and Archbold 1945, McTainsh and Walker 1982) from where the dust is transported during the months of November to March in a WSW direction out over West Africa and the Atlantic Ocean.

DUST DEPOSITION PROCESSES AND PARTICLE-SIZE

Dust deposits collected at different locations and times during three Harmattan seasons (1976-1979) in northern Nigeria, cover a size range of 2 to 200μm (9 to 2.25ø), although all samples except one, had 50% by weight in the 2 to 63μm (9 to 4ø) fraction. This size range may reflect both sediment supply and deposition processes, however the attention in the present paper is upon the latter.

A number of aeolian processes operate during dust deposition and their effects upon the particle-size of the deposited dust will be considered here.

Spatial patterns of dust deposit particle-size

As dust settles out of the Harmattan Wind over northern Nigeria a spatial pattern of deposit particle-size emerges, which can be expressed in two dimensions. Firstly, there is a clear pattern of decreasing dust deposit particle-size with distance downwind (Fig.2; locations Fig.1) which reflects the preferential deposition of coarser particles close to source, with increasingly fine dust deposition with distance downwind. Over the Western Atlantic Ocean dust particle-sizes become significantly smaller. Prospero et al (1981) report 91.5% of material <12μm (6.4ø). The second dimension of the spatial pattern is the dust deposit particle-size decrease normal to the dominant wind direction (Fig.3; locations Fig.1). The same two dimensional spatial pattern exists with dust deposition rates (McTainsh and Walker 1982).

These spatial patterns of dust particle-size may reflect two characteristics of the Harmattan Aeolian System. Firstly, the Harmattan wind is relatively consistent in terms of wind direction. Data from Kano over a 13 year period show that 87% of the winds during the Harmattan season blow from between north and east. Secondly, Harmattan dust appears to radiate downwind from a point source, the Bodele Depression, which may explain the two dimensional pattern of particle-size. By contrast, the Illinois loess apparently eminates from a linear source, the Mississippi River floodplain (Smith 1942).

Spatial patterns of loess soil particle-size have been identified in the USA by Smith (1942) and Frazee et al (1970), and in China by Qian Ning et al (1980), and an aeolian size-selection process inferred. Until the present study, however, little actual process evidence has been presented to give validity to such inferences. The two dimensional pattern described here may be cause for caution in using spatial trends in loess soil particle-size to infer paleo-wind direction or strength. In addition, the range of dust deposit modal particle-size is from 90μm to 12μm, depending on location, which exceeds the 20-50μm loess criterion. The 90μm mode is also in excess of the 70-80μm dividing size between suspension and saltation particles proposed by Bagnold (1941).

Temporal variations in dust deposit particle-size

A basic premise behind the interpretation of the spatial pattern of dust particle-size is that under the same wind conditions coarse particles will be transported and deposited over a shorter distance from the dust source area than finer dust particles. In reality, however, wind conditions fluctuate considerably throughout a particular season. Fig.4 demonstrates two aspects of temporal changes in dust particle-size. Firstly, the particle-sizes of the dust deposited at Kano and Zaria do change quite dramatically from week to week. Secondly; there appears to be a general correspondence between the character of deposits laid down simultaneously at the two locations (Fig.1). The November 24 - December 1, 1978 deposits at Kano and Zaria are better sorted and coarser than the December 29 - January 5, 1979 deposits at both sites. Also, the two remaining deposits are generally poorly sorted.

There appears to be an overall control on dust deposit particle-size at these two sites, which is probably the windspeed (or competence) of the depositing wind. Fig.5 demonstrates that as wind speed at Kano increases, so does the particle-size of the deposited dust. In addition, McTainsh and Walker (1982) have identified that diurnal changes in wind conditions also affect dust deposit particle-size.

In summary, it is apparent that dust particle-size is sensitive to changing wind conditions, even over the relatively limited time scale of the present measurements. If these relationships are viewed within the broader time context required for the formation of a loess deposit, it appears reasonable to assume that changes in wind conditions alone would produce a loess deposit with a particle-size range which exceeds the 20-50μm criterion.

CONCLUDING COMMENTS

The dust transported and deposited from aeolian suspension in the Harmattan wind exhibits a greater particle-size range than the 20-50μm range assumed to represent all loess deposits. Also, this dust particle-size diversity is explicable in terms of such processes as transport distance and wind conditions.

In conclusion, it is proposed that, although the 20-50μm loess size criterion may be a realistic for glacial loess deposition processes, it is unlikely that desert loess deposits of such a narrow size range could be produced by Harmattan dust deposition processes. It follows therefore, that the relative scarcity of undisputed desert loess deposits, possibly reflects more upon our limited understanding of desert loess deposition processes, than on the existence or not of desert loess.

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Figure Captions

Fig. 1 Northern Nigeria - Study Area

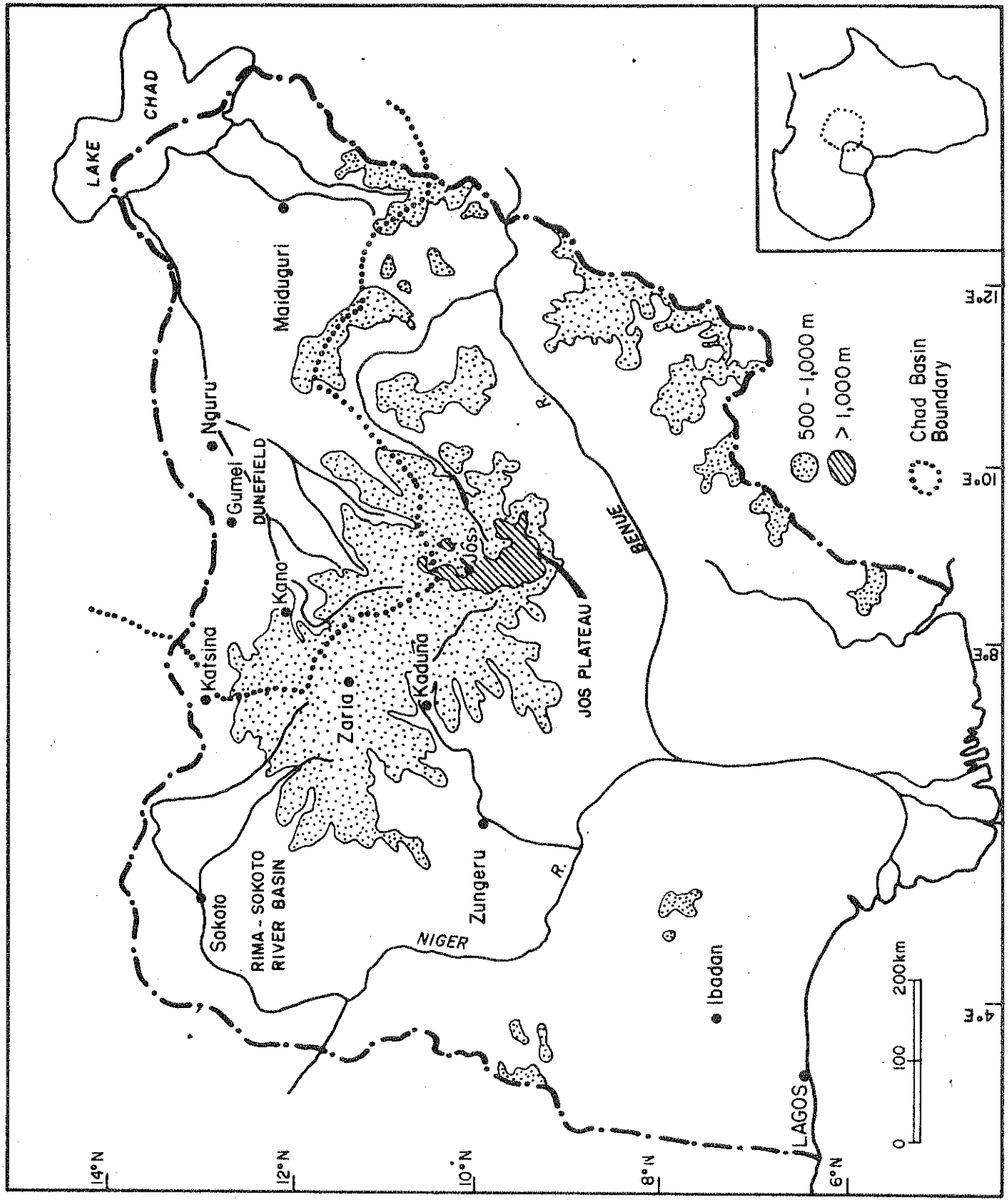
Fig. 2 Relationship of Harmattan dust particle-size to distance downwind from source area.
(after McTainsh and Walker 1982)

Fig. 3 Relationship of Harmattan dust particle-size to distance normal to the predominant wind direction.
(after McTainsh and Walker 1982)

Fig. 4 Temporal changes in dust particle-size at Kano and Zaria.

Fig. 5 Relationship between dust deposit particle-size and windspeed - January 20-25th 1979, Kano.

Fig. 1


LAKE CHAD
Maiduguri
Nguru
Gumel
DUNEFIELD
Kano
Katsina
Zaria
Kaduna
Jos
JOS PLATEAU
BENUE
R.
Zungeru
R.
NIGER
Sokoto
RIMA - SOKOTO RIVER BASIN
Ibadan
LAGOS
500 - 1,000 m
> 1,000 m
Chad Basin Boundary
0
100
200km
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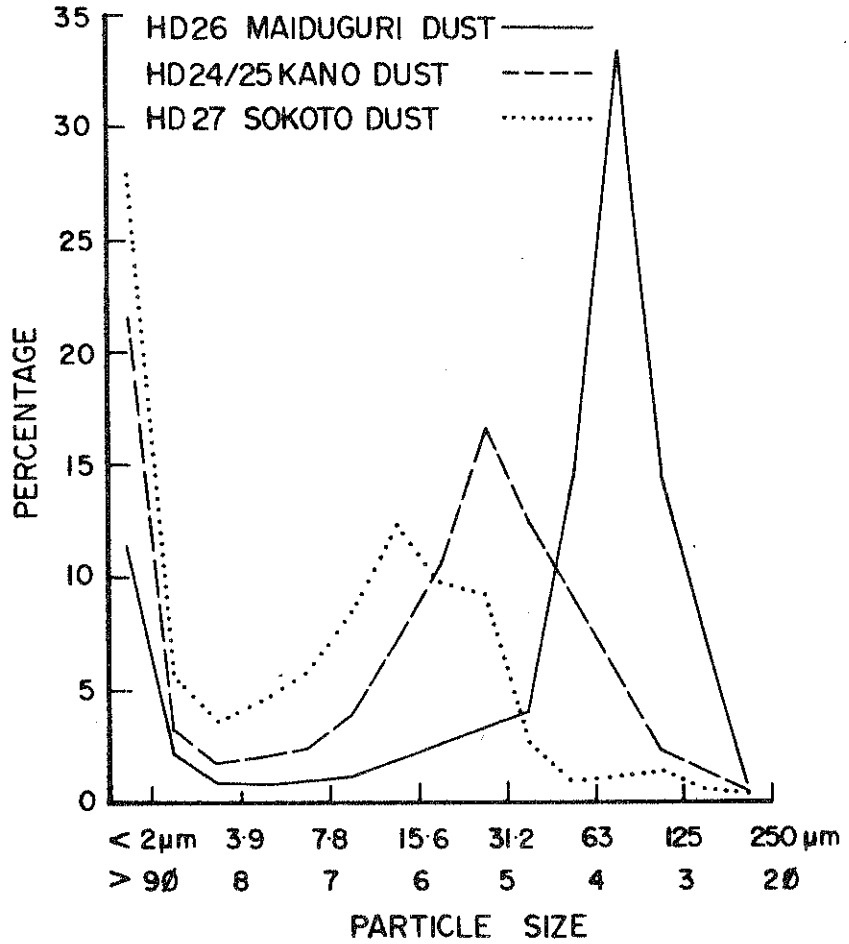
10°N
8°N
6°N
4°E
8°E
10°E
12°E

Fig. 2.


HD26 MAIDUGURI DUST
HD24/25 KANO DUST
HD 27 SOKOTO DUST
PERCENTAGE
35
30
25
20
15
10
5
0
< 2µm
3·9
7·8
15·6
31·2
63
125
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> 9Ø
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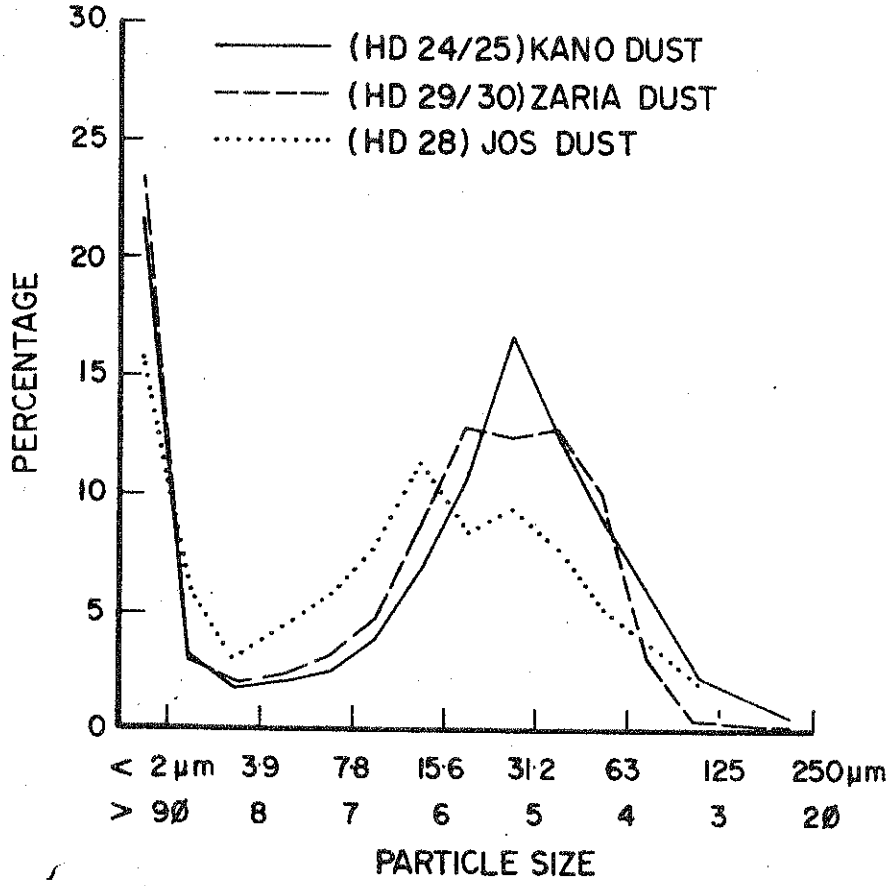
PARTICLE SIZE

Fig. 3


(HD 24/25) KANO DUST
(HD 29/30) ZARIA DUST
(HD 28) JOS DUST
PERCENTAGE
30
25
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0
< 2 µm
3·9
7·8
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31·2
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125
250µm
> 9Ø
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3
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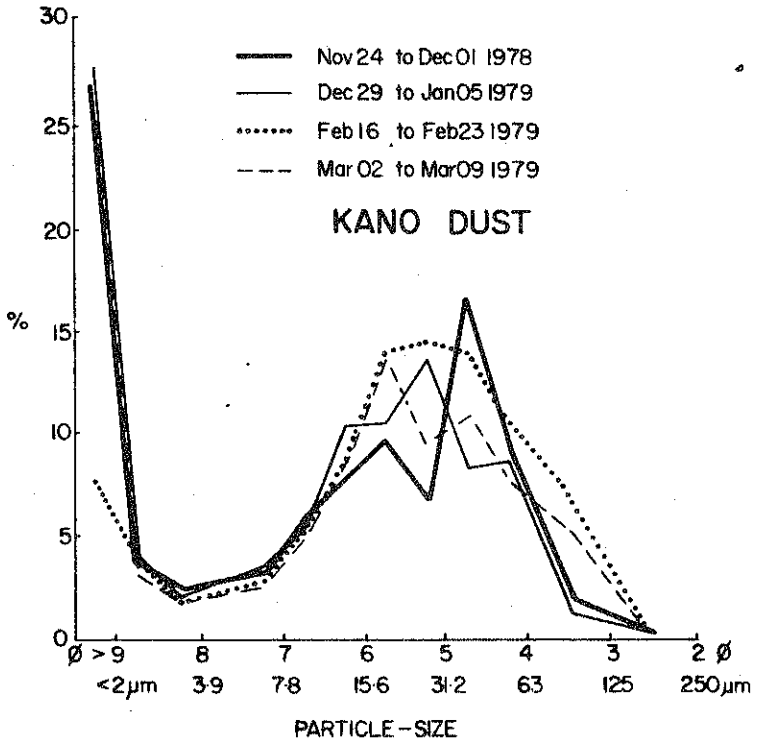
PARTICLE SIZE

Fig. 4


Nov 24 to Dec 01 1978
Dec 29 to Jan 05 1979
Feb 16 to Feb 23 1979
Mar 02 to Mar 09 1979
KANO DUST
%
30
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8
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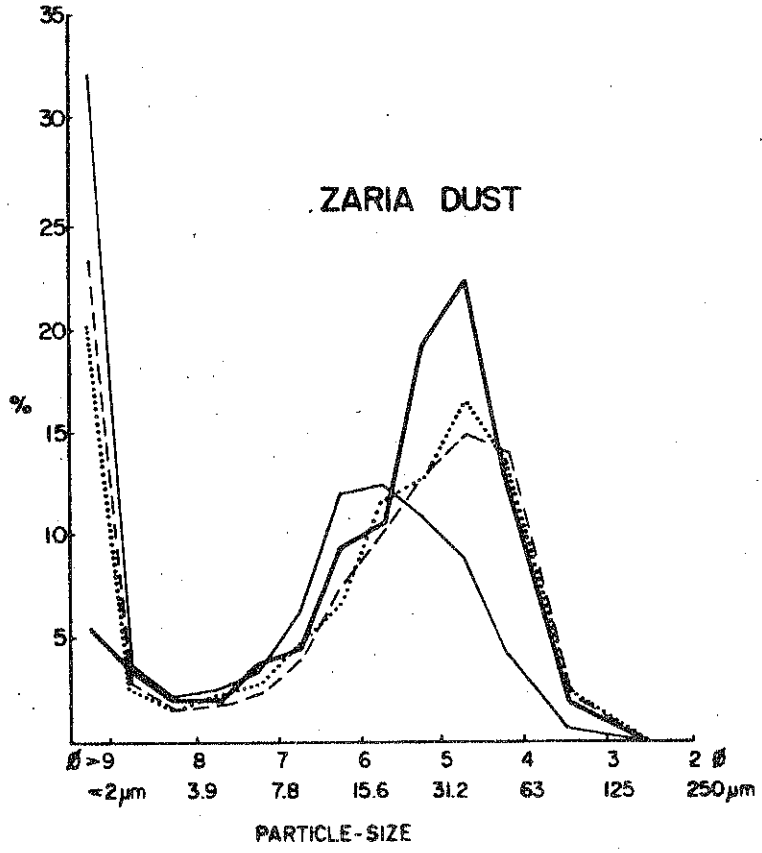
PARTICLE-SIZE


ZARIA DUST
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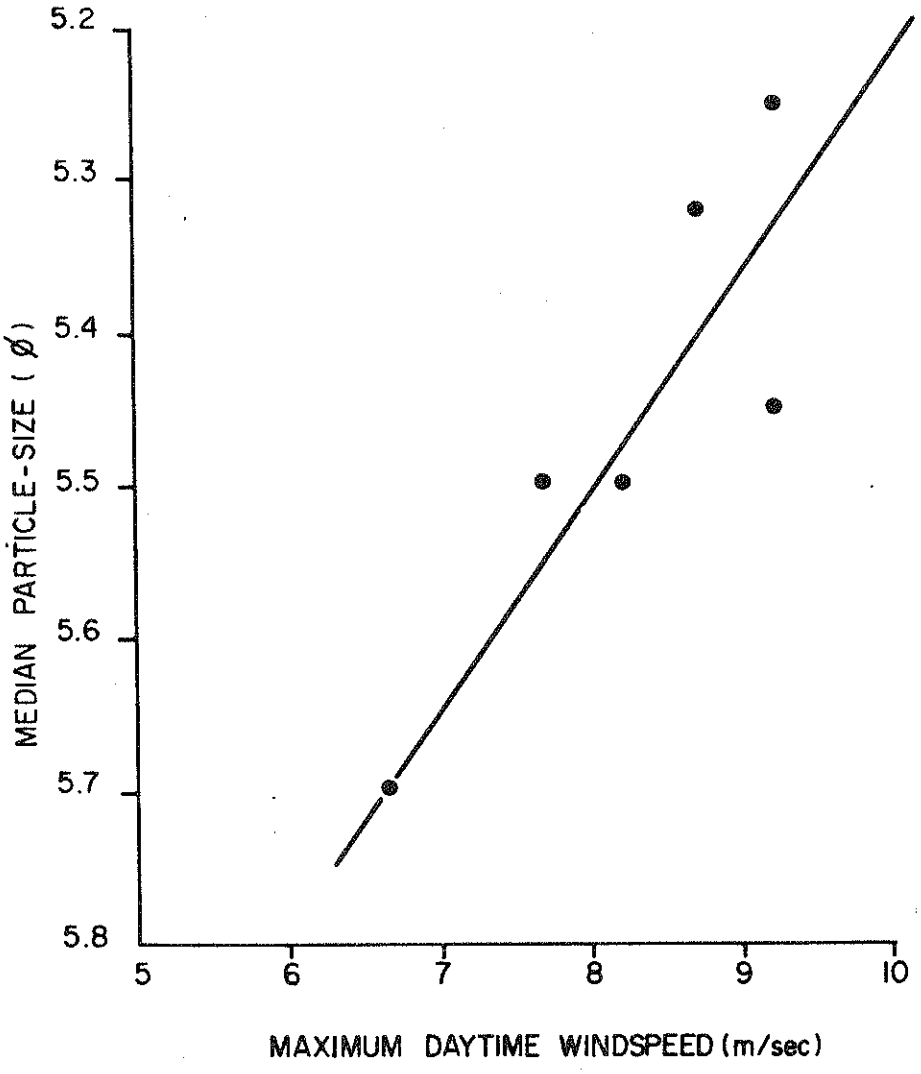
31.2
63
125
250μm
PARTICLE-SIZE

Fig. 5


5.2
5.3
5.4
5.5
5.6
5.7
5.8
MEDIAN PARTICLE-SIZE (ø)
5
6
7
8
9
10
MAXIMUM DAYTIME WINDSPEED (m/sec)

A STATISTICAL ANALYSIS OF CHANNEL MIGRATION WITH IMPLICATIONS FOR DETERMINING BEDLOAD TRANSPORT RATES IN MEANDERING RIVERS

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Abstract

Mean migration rates for 18 meandering river channels in Western Canada are explained statistically in terms of hydraulic and sedimentological variables. The volume of sediment eroded from the outside bed and bank of a meander bend is shown to be largely a function of river scale and the grain size of sediment at the base of the floodplain; together they explain statistically nearly 70% of the volumetric migration rate. Channel migration is essentially a problem of sediment entrainment dependent on total stream power and sediment size, the same variables that determine the rate of bedload transport. The down-valley sweep of a bend in a meandering river results from the erosion of the concave bed and bank and the nearly equivalent deposition of sediment on the convex bed and bank. This process of channel shifting gives rise to a definite average transport rate of eroded material. Bedload transport is shown to be directly equivalent to, and accurately predicted from, measured rates of channel migration.

Introduction

The objective of this paper is to statistically relate bend migration rates for 18 river reaches to channel hydraulic, geomorphologic and sedimentologic characteristics in an attempt to develop a predictive, empirical model of channel migration for single-thread meandering rivers in Western Canada. Furthermore, we demonstrate that lateral migration is largely determined by the same variables that control sediment entrainment and transport. As a consequence, we show that bedload transport can be relatively simply determined from lateral migration rates of meandering rivers.

In earlier work (see Hickin & Nanson, in press) we have suggested that the rate of channel migration (M) is likely dependent on stream power per unit area of the bed (ω), channel width (W), the force per unit area of the outer (concave) bank resisting channel migration (τ_b), the bank height (h) and the bend radius of curvature (r).

$$M = f(\omega, W, \tau_b, h, r,) \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ldots 1$$

Study Rivers and Data Determination

The eighteen river reaches examined here are located in British Columbia and Alberta (Fig. 1). They are single-thread, meandering, gravel and sand-bed rivers for which sequential aerial photography provides evidence of channel shifting during the last 21-33 years. Discharges at the 5-year flood vary from 44 to 3972m^3s^{-1}, channel slopes from 0.0016 and 0.00022, sediments from fine sand to large cobbles, and migration rates from zero to 12.6 my^{-1}.

Individual measurements of channel migration, channel width and radius of bend curvature were obtained for 118 freely meandering bends of the 18 river reaches. In addition, average slope (S) height of the concave outer bank, median diameter of basal sediments in the outer bank (D_{50}) and magnitude of the 5-year flood ($Q_{5.0}$) were obtained for each study reach following the methods described in Hickin and Nanson (in press).

Channel planform data were measured from maps drawn from screen images projected by 35mm transparencies of British Columbian, Albertan and Canadian Government 1:15000 to 1:40000 aerial photographs. Channel displacement was measured by superimposing two images of the same channel reach time lapsed by 21 to 33 years. Common registration points were used to match each pair of images.

Data Selection

Our previous research on the Beatton River has clearly shown that channel migration is a complex function of bend curvature (Hickin and Nanson, 1975; Nanson and Hickin, 1983) (Fig. 2). Similar data presented here for a total of eighteen rivers are enveloped by a curve displaying the same basic form as the Beatton River data, with a maximum migration rate occurring at $2.0 < r/W < 3.0$ (Fig. 3).

In order to isolate the effect of bend curvature on channel migration, a stratified sample of bends were selected with $2.0 < r/w < 4.0$, such that curvature could be regarded as being held essentially constant. Because of the intermittent nature of channel migration, bends which showed zero migration were omitted from the analysis (see Hickin and Nanson, in press). This stratified data set provided a mean optimum migration rate for each river.

Data Analyses

In order to establish the respective variance contributions and hence the predictive value of each independent variable (and their derived parameters) from Equation (1), a series of stepwise regressions were executed. To test for statistical relationships among the variables in Equation (1) the data were grouped by river reach, and simple means derived for migration rate (M, my^{-1}) and channel width (W,m). These, along with discharge for the 5 year flood ($Q_{.5.0}$, m^3s^{-1}), channel slope (S), outer bank height (h, m) and the median diameter of the basal sediments in the outer banks (D_{50},mm) were used to derive the parameters of bank erosion rate as a volume per unit length of channel per year (Mh, $m^3m^{-1}\ y^{-1}$: $m^2\ y^{-1}$), power per unit area of the bed (ω, $\rho gQSW^{-1}$: WM^{-2}) and total power per unit length of channel (Ω, ρgQS: Wm^{-1}). Because all these sets of data are strongly positively skewed towards the largest rivers, their distributions were normalized by log transformation.

The simple correlation matrix of transformed variables (not presented here) indicates that no single variable accounts for more than 50% of the variance in the mean migration rate, with total stream power, mean width, and discharge providing 48%, 44% and 34% respectively. Using stepwise regression, mean migration rate is regressed sequentially against groups of independent variables introduced in order of their decreasing simple correlation. With this procedure, additional independent variables were added to the relationship only if, in combination, they explained more than 5% of the variance in mean migration, and then only if their individual contribution was significant at at least the 10% level.

By comparison with regressions conducted with only migration rate (M) as the dependent variable (not presented here) we have shown that the use of the volumetric migration rate (Mh) gives a much better prediction of channel migration from measurements of channel width, discharge, slope, bank height and basal sediment size.

$$Mh = 25.06Q^{0.788}S^{0.74}D_{50}^{-0.209} \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad 2$$

69.1	57.5	2.6	9.0	(% explained variance)
1.0	1.0	2.5	5.0	(% signif. of explained variance; F test)
	1.0	5.0	10.0	(% signif. that exponent ≠ 0; T test)

$$Mh = 2.089W^{1.369}D_{50}^{-0211}S^{0.568} \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad 3$$

62.6	53.6	1.5	7.4
1.0	1.0	10.0	10.0
	0.1	10.0	10.0

$$Mh = 41.975\Omega^{0.795}D_{50}^{-0.215} \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad 4$$

68.2	42.1	26.1
1.0	1.0	1.0
	0.1	1.0

$$Mh = 0.607\omega^{0.823}D_{50}^{-0.207} \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad 5$$

29.3	11.5	17.8
10.0	2.5	10.0
	5.0	10.0

Equations (2), (3) and (4) show that the scale of the river (expressed as discharge or width) and river slope provide important statistical explanations of migration rate. The fact that width seems to be nearly as good a predictor of volumetric migration rate as does discharge probably indicates that width integrates the effect of the long-term flow record to nearly the same extent as do the relatively short-term discharge records used to calculate $Q_{5.0}$. This must be of some comfort to those wishing to estimate bend migration rates on ungauged rivers.

When discharge and slope are combined in the physically meaningful parameter of stream power, then it alone explains only 42% of the migration rate variance, but grain size jumps in importance to contribute a further 26% (Equation 4). However, stream power per unit width is a poor predictor of channel migration rate explaining only 11.5% of the variance, with grain size contributing a further 18% (Equation 5).

Equations 2 and 4 provide the best predictive capability of the above set although the unexplained variance remains high at approximately 30% in both cases.

They show that river size is the most important contributor to channel migration. However, if volumetric erosion rate (Mh) were to be scaled by stream power, it could provide a basis for examining the relatively small influence of basal sediments on the rate of lateral erosion, using the equation

$$\frac{Mh}{\Omega} = kD^{a}{}_{50}$$

Furthermore, if the dependent variable were to be inverted, it could then be interpreted as a measure of bank resistance (Equation 6), a parameter which infact has the dimensions of shear stress (force/area; Nm^{-2}) (Hickin and Nanson, in press).

$$\frac{\Omega}{Mh} = 185.78D_{50}{}^{0.265} \qquad \ldots\ldots\ldots\ldots\ldots\ldots\ldots \quad 6$$

61.1	61.1
1.0	1.0
	0.1

where $\frac{\Omega}{Mh} = \tau_b$

Equation 6 shows a strongly positive statistical relation between the size of basal sediments and bank resistance (τ_b) for basal sediments coarser than silt (Figure 4). Although τ_b has the dimensions of force/area, it is a coefficient of resistance to lateral migration, presumably dependent largely on bank strength, but also absorbing all the other factors including the statistical variability in stream power and migration rates.

The relationship between channel migration and bedload transport

Discharge, power or even channel width explain more than 42% of the variance in volumetric erosion of the outer bank, and in combination with measurements of sediment size at the base of the outer bank, they explain up to 70% of this variance. Holding river scale constant, these results show the size of basal sediment in the outer bank to be particularly influential in determining erosion rate (Equations 6). Surprisingly, bank vegetation appears to be a relatively unimportant influence on the rate of migration on medium to large rivers. Numerous cross-sections surveyed in meander bends show that lateral channel migration involves substantial erosion of the laterally sloping bed in the outer part of the

channel, as well as erosion of the near vertical and possibly well vegetated upper part of the bank (also see Thorne and Lewin, 1979: Thorne and Tovey, 1981). The use of a substantial vertical exaggeration in constructing most channel cross sections has over emphasized the dominance of the banks and the extent of bank vegetation within these sections. Indeed, during lateral migration bed erosion may be required to move up to half the vertical extent of floodplain sediment. Once the outer bank is undermined by this process, even well vegetated and cohesive upper sediments will collapse into the river and disintegrate. In other words, bank erosion and channel migration are largely determined by bed-material transport. It is for this reason that a simple statement involving stream power and basal sediment size provides such an effective means of expressing the driving and resisting forces in this predictive model of channel migration. The model properly addresses a bedload transport process (Bagnold, 1980) rather than one involving bank shear as assumed by Begin (1981).

Strong support for the argument that channel migration and bedload transport are causally related, and indeed, that they are very closely correlated, has been provided by Neill (1971, 1984). He has shown theoretically (1971) that channel migration in an equilibrium system must give rise to a definite rate of transport of the eroded material, such that:

$$Q_s = l_t h M$$

where Q_s is the volumetric transport rate, l_t is the average travel distance of sediment between erosion and deposition (approximated as the arc length of a half meander wave), and M is the average migration rate (in the same time units as Q_s). Somewhat later, Neill (1984) obtained reliable bedload sampling data from Burrows et al (1981) who used a Helley-Smith sampler for a 5 year period (1977-81) on the Tanna River in Alaska. They obtained annual yields averaging 360,000 tons of bedload per year. Calculations of bedload obtained from very detailed measurements of channel migration in a bend adjacent to the sampling site provided an average annual bedload yield of 334,000 tons. While one swallow does not make a summer, the remarkable similarity between these two figures is very encouraging. Channel migration is a relatively simple function of stream power and the size of basal sediment in the channel and floodplain. Predicting channel migration on meandering rivers appears to be a sediment transport problem, and vice versa.

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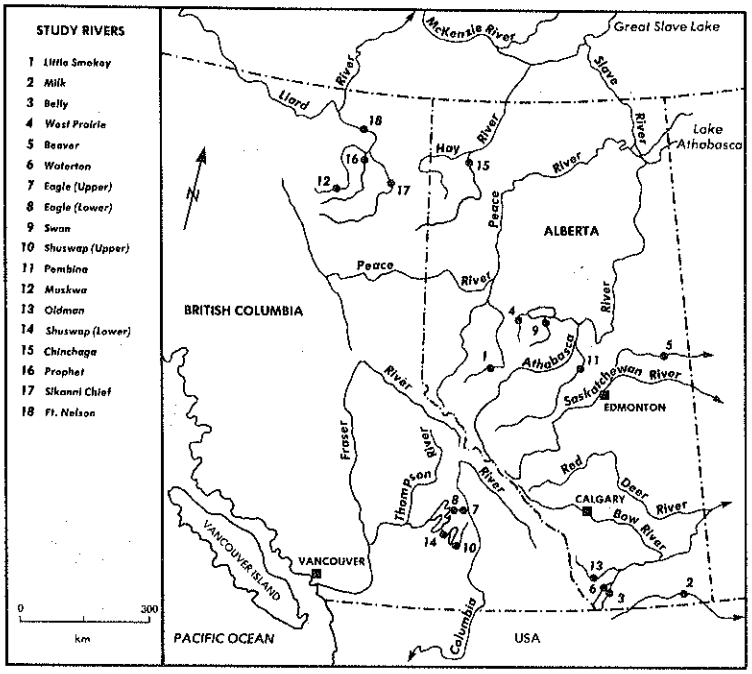

Figure 1 Study Rivers

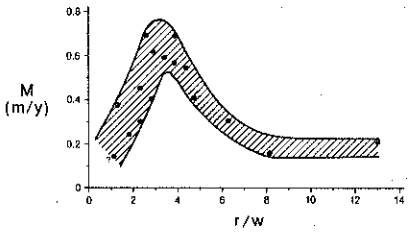

Figure 2 The relationship between migration rate (**M**) and bend curvature(r/w) for the Beatton River.

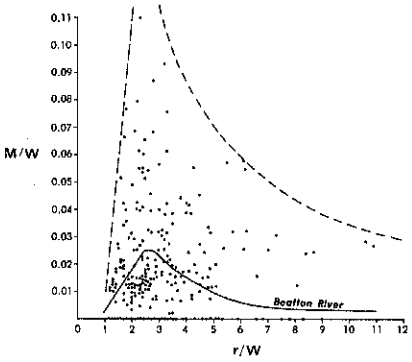

Figure 3 The relationship between relative migration rate (units of channel width) and bend curvature for all field sites.

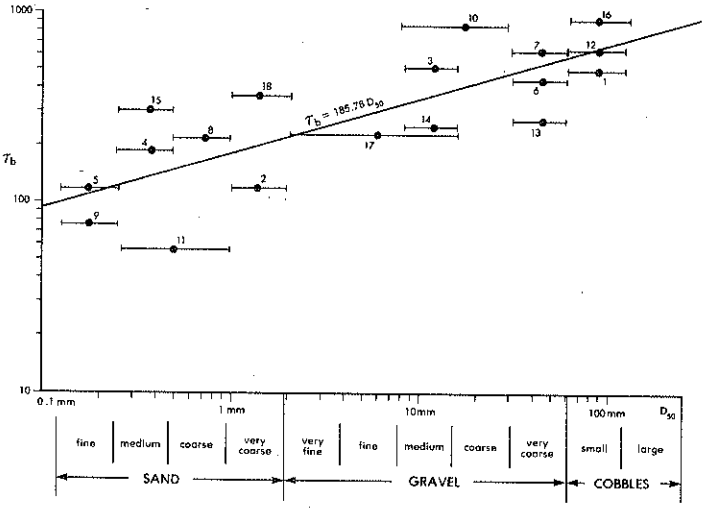

Figure 4 The relationship between the coefficient of resistance to lateral erosion and the median diameter of the basal sediments. Horizontal bars define the Wentworth textural class of the grain size of the basal sediment at each study reach.

SEDIMENT DEPOSITION IN LAKE WYANGALA NEW SOUTH WALES

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ABSTRACT

Information on Lake Wyangala sediment deposits obtained during a period of low storage levels provides clues to the nature of several depositional processes. A thin and uniform fine-textured veneer over the main basin is deposited by still-water settling of clay and organic matter. A downstream-fining deposit in tributary arms is deposited by the settling of larger particles when inflows meet the reservoir backwater. Interbedded and laminated sands, silts and clays are deposited along the entire length of pre-dam channels by a combination of low water level inflows, density currents and still-water settling. This suite of deposits is different from that usually found in lakes or estuaries due to the great variability of water levels in this irrigation reservoir.

INTRODUCTION

Lake Wyangala's storage capacity of 1.2 X $10^9 m^3$ is used for irrigation along the Lachlan River. The recent drought in New South Wales resulted in very low storage levels in Lake Wyangala. The combination of low rainfall in its catchment of 8300 km^2 and a high demand for irrigation water resulted in storage volumes less than 10% of maximum for periods of several months from 1980 to 1983. These low levels exposed bottom sediments, allowing the opportunity for studying their location, thickness and properties to determine total sediment volume, sediment sources and deposition processes. Information on sediment deposition is an important contribution to the study of the drainage basin sediment system and sources (Outhet, 1984). It is also important when determining erosion history from sedimentation (Wasson et al, 1984).

METHODS

The first activity of the study was the acquisition of high quality, large scale (1:10,000) colour air photography which showed the location of all exposed sediment in February 1981 when the storage was at 14% of full capacity. These photographs were used to locate deposits and choose 50 study sites.

The thickness of exposed sediment is usually determined by digging a pit down to the pre-dam soil surface. These pits also reveal the presence of laminations, whether the sediment is uniformly textured or interbedded, and allow the collection of samples from specific layers. A sediment probing rod is also used.

Determining the thickness of submerged sediment is more difficult. In Lake Wyangala, this has been done by estimates from silt section resurveys. However, this method is inaccurate (0.5 m resolution) because of low technology surveying, the practical problem of positioning and inconsistency of methods. To overcome this, a seismic profiling technique has been developed by this author and

Mr. J. Tayton, a geophysicist at Macquarie University. This technique uses sound refraction and reflection to determine the sediment/soil interface and thereby measure the sediment thickness directly and immediately. The resolution of this method is 0.2 m.

RESULTS

Early data gathered from the air photographs, maps and initial field work in 1981 revealed three main types of deposit. The following sections present more information about these three deposits. Much of this information is shown in Figures 1 and 2. Figure 3 shows a preliminary map of the sediment deposits in a part of Lake Wyangala.

Main Basin Deposit

One of the pre-dam terraces found in several parts of the main basin is a representative site of a typical main basin deposit. It is located well away from the pre-dam channel and side slopes. Observations in this area were made at 260 points located at approximately 5 metre intervals along 3 transect lines. The mean sediment thickness is 150 mm with a standard deviation of 30 mm. The material is faintly colour laminated (within Munsell group 1) with a uniform medium clay texture at all sites. Similar sediment properties and thicknesses were observed at all of the 10 other study sites in the main basin.

Deposits on the reservoir side slopes thin out as elevation increases (Figure 1). This relation was found at all sites in Lake Wyangala. The elevation at which sediment thickness thins to zero is called the "zero line" and is found at approximately the same elevation at all sites in the main basin of the reservoir (351 m). This elevation may be related to duration of submergence and erosion on the slope above the zero line. This relationship is under investigation.

Tributary Arm Deposit

In an upstream direction away from the main basin deposit along either the Lachlan or Abercrombie arms, surveys show that the zero line slopes upward, the sediment texture becomes coarser and colour becomes lighter (Munsell group 2). Colour laminations become more distinct. This is called the tributary arm deposit. Figure 2 shows a profile of this deposit on the Abercrombie River Arm and includes observations from 25 sites. Sediment texture is from field texturing and laboratory analysis.

Channel Deposit

Wyangala Dam is operated for irrigation purposes. Accordingly, the stored water volume is highly variable and can be reduced from 100% of storage capacity to 10% in one year if catchment rainfall is low and water demand is high. For this reason, the upstream limit of backwater in the pre-dam channels can vary by 22 km in distance and by 40 m in elevation (Figure 2).

The pre-dam channels of the Abercrombie and Lachlan Rivers contain a deposit quite different from the two previous ones. It is made up of many layers of contrasting texture which are interbedded clays, silts and sands. This deposit is much thicker than the others (up

to 7 m). It thins near the dam wall and near the upstream backwater limit. There are no deltas at the upstream limits of dam backwater on the tributary arms.

PROCESSES

The above information leads to several conclusions about the processes that account for the three types of deposit.

The uniform and fine-textured nature of the main basin deposit indicates that it is formed by fine particles settling slowly out of still water.

The vertically uniform but downstream-fining texture of the tributary arm deposit indicates that it is formed by particles settling out of moving water with decreasing velocity and turbulence (competence) in a downstream direction.

The interbedded and laminated channel deposit is formed in a highly variable depositional environment by both moving and still water with variable sediment loads. This variation is due to a combination of the two processes mentioned above in addition to two others. One is inflows moving along exposed pre-dam channels during low reservoir levels. The second is occasional density (underflow) or turbidity currents moving along submerged pre-dam channels during high water levels. One density current was measured at Burrinjuck Dam. It occurred in March 1983 and contained almost 24% solids when sampled just upstream of the dam wall. Density currents are formed when an inflow has a higher sediment concentration, higher salinity or a lower temperature than the reservoir water.

A DEPOSITIONAL MODEL

All the preceding information forms the basis for a model of the depositional system in Lake Wyangala. The system receives almost all its original sediment input as tributary inflows laden with gravel, sand, silt, clay and organic matter. As it travels down a tributary channel towards the reservoir backwater, an inflow may rework any channel sediment deposited by previous inflows at higher backwater levels. When it reaches the backwater, gravel and sand in the bed load settles out immediately over any previously-deposited sediment to form the channel deposit. As the inflow plume proceeds into the backwater zone it creates turbulent currents in the tributary arm which can carry fine sand and silt to elevations above the pre-dam channel (flood plains, terraces and side slopes) to form the tributary arm deposit. Finally, the clay and buoyant organic matter is carried by slow currents caused by any of the many hydraulic, wind or thermal forces (Humphries and Imberger, 1982) to the distant main basin. When the turbulence ceases or when particles settle below the turbulent surface zone they accumulate on the bottom to form the main basin deposit.

The model described above is different from that which produces the well-known topset, foreset and bottomset bed system found in lakes and estuaries as first described by Gilbert (1890). It has been modified by the great variability of backwater level for each inflow. Part of the channel deposit and the tributary arm deposit are foreset and topset beds that have been distributed along tributary arms. Only the constantly-submerged part of the main basin deposit is

similar to the usual bottomset beds.

MINOR PROCESSES

Several subaerial processes have been observed to have minor effects on the three types of sediment deposit. These processes begin to modify any deposits as soon as they are exposed by low water levels. In the pre-dam channels, inflows often form knickpoints and various bed forms. On side slopes, wave action at the shoreline makes small erosion scarps and beaches composed of clay granules at low elevations or sand at the higher elevations. Some of the fine particles sorted from the material on the side slopes may become mixed into the ponded water and then deposited on the bottom. Bio-deposition after algae blooms, for example, may add diatoms to bottom sediments. The exposed side slopes are sites of sheet, rill and gully erosion with the extent of each determined by the site's elevation and, consequently, duration of exposure. All of these processes have the effect of moving the sediment downstream or downslope to the lower regions of the reservoir and concentrate it there.

CONCLUSION

In a lake or estuary, the major sedimentation processes mentioned in this paper would produce the usual topset, foreset and bottomset bed sequence of deposits. However, the information gathered in this study has revealed that large water level variations produce a very different set of deposits, namely a main basin deposit, a tributary arm deposit and a channel deposit. In addition, these deposits are modified by subaerial erosion processes that tend to concentrate the sediment in the lower regions of the reservoir.

ACKNOWLEDGEMENTS

The author gratefully acknowledges assistance from the New South Wales Water Resources Commission and comments from Dr. R. Blong of Macquarie University, School of Earth Sciences.

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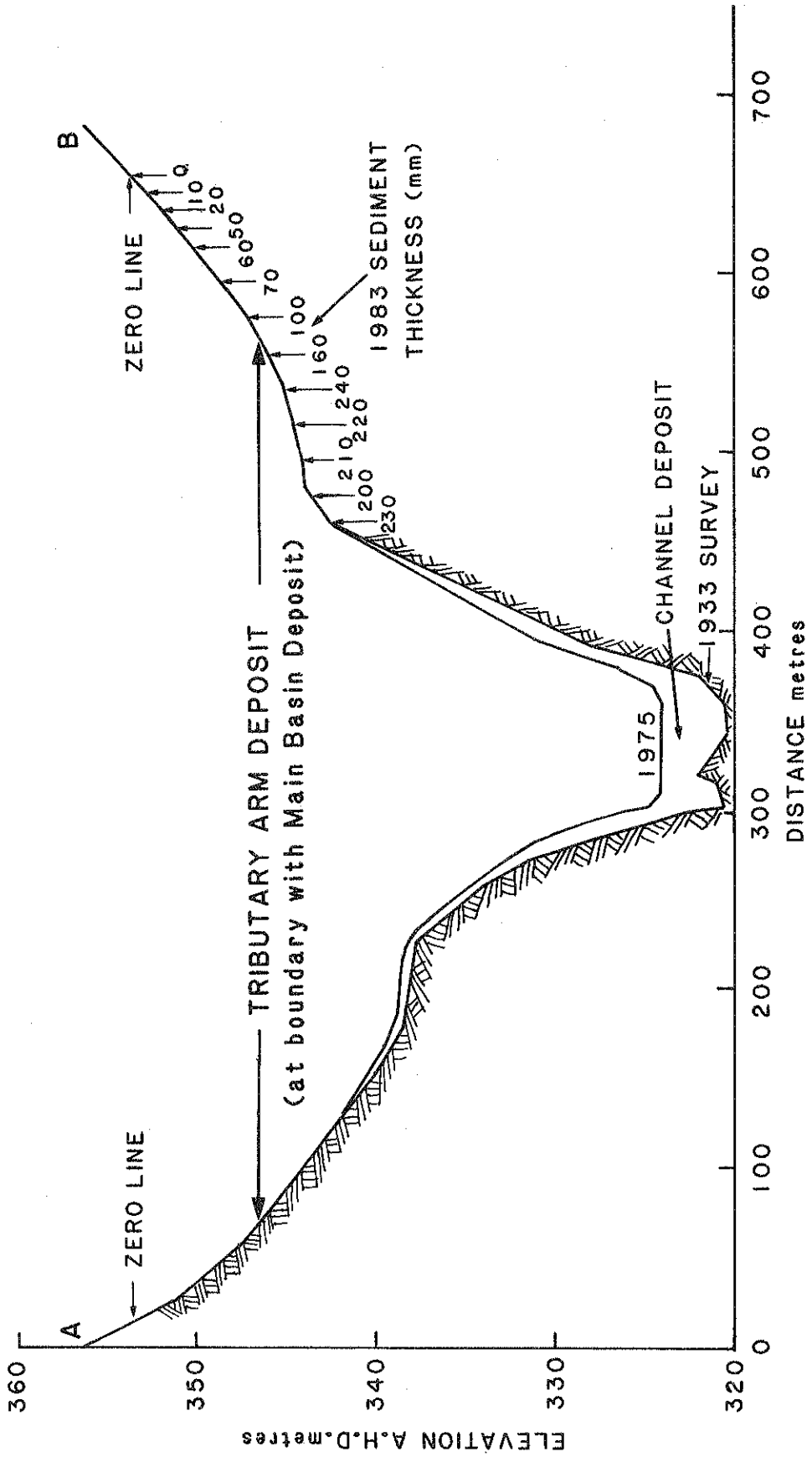

Figure 1. Cross Section, Lachlan River Arm. (See Figure 3 for location).

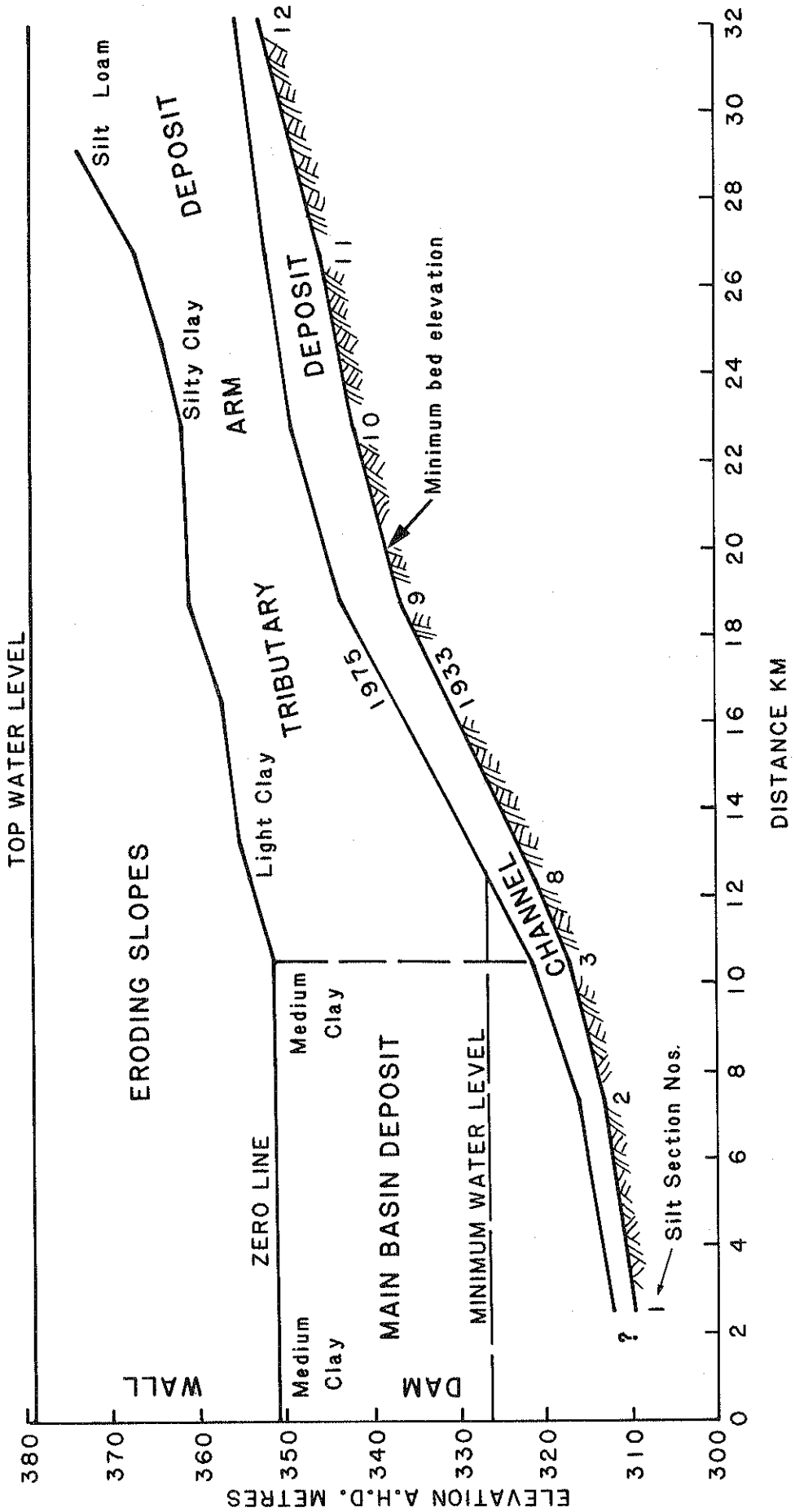

Figure 2. Profile of Abercrombie River and Lachlan River upstream from Dam Wall.

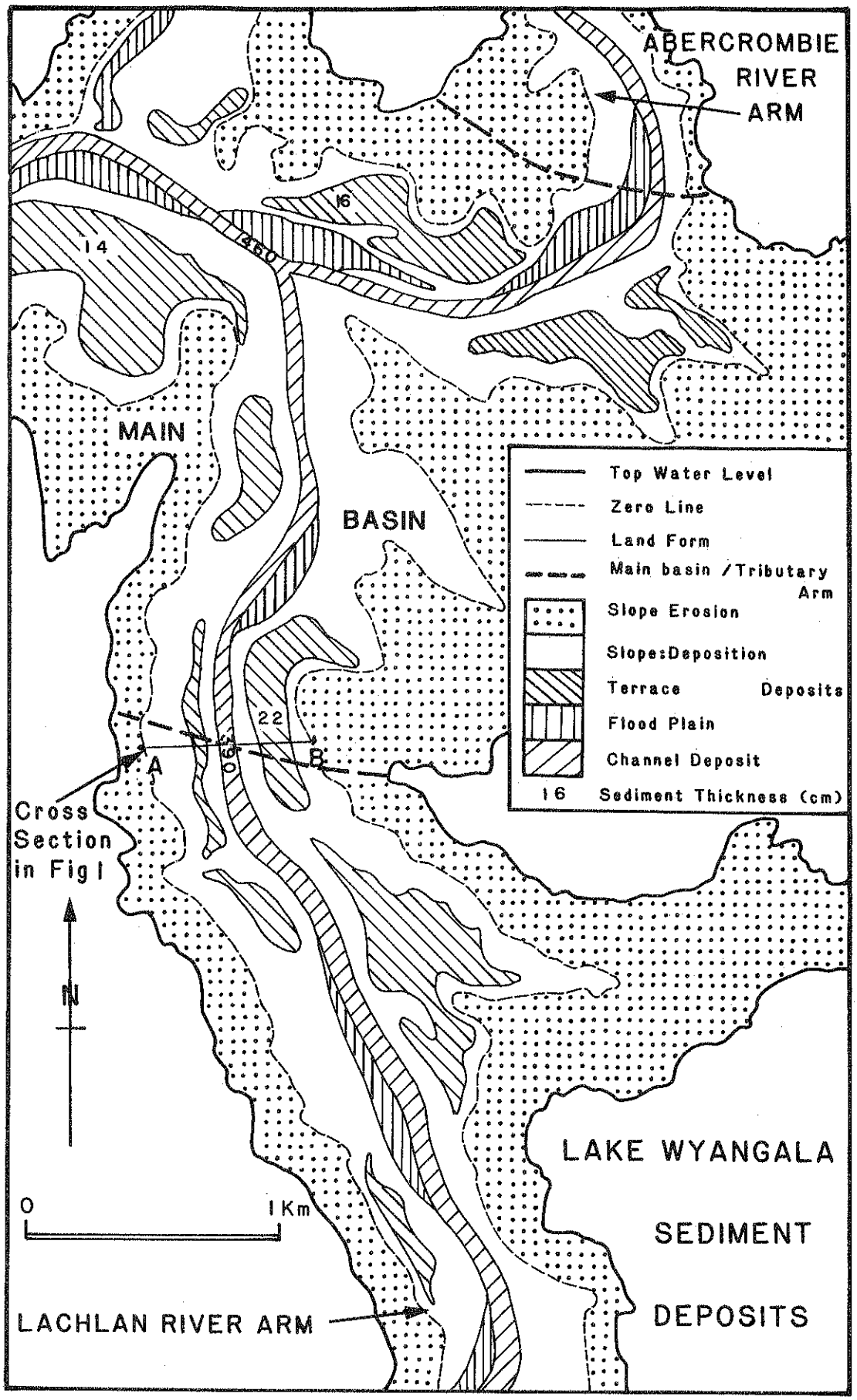

Figure 3. A Part of the Sediment Deposit Map for Lake Wyangala.

A.N.Z. Geomorph. Group, 2nd Conf., Broken Hill, July 1984.

LANDSAT IMAGERY AND LANDFORM MAPPING

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ABSTRACT

Published landform classification are not easily used with Landsat imagery. A simple classification using patterns (especially lines), texture and shape is presented. For broad scale landform mapping, Landsat imagery in better than either aerial photographs or topographic maps.

INTRODUCTION

Landscape and landform mapping are basic to geomorphology. Traditionally such mapping has been done from topographic maps or aerial photographs. However, with the advent of orbiting satellite platforms for remote sensing systems, new kinds of imagery, such as synthetic aperture radar, thermal infrared images, and the multispectral images of Landsat 1-4, are becoming increasingly available. Using Landsat imagery, this paper addresses the questions of techniques of geomorphic mapping from Landsat images, and whether new classification ideas are needed.

At first sight, landform mapping from Landsat images seems like aerial photograph interpretation, only harder, because of the lack of stereoscopic viewing allowed by the latter. However, there have been few attempts to develop methodologies of landform mapping that incorporate Landsat images, and even fewer attempts to use remote sensing in geomorphic work in Australia.

The basic idea behind any landform classification is that all landscapes can be divided into smaller units. The method by which this division, or classification, is carried out will depend on such things as scale, purpose, and the kind of data to be used. Because Landsat images cover a large area (185 km x 185 km), and have a ground resolution of about 80 m, only broad scale landform classifications are considered here.

Examples of landform classification developed in Australia are those of Loffler and Ruxton (1969), Mabbutt (1973), and Loffler (1974). However, these classifications were not developed for Landsat images, and it would be surprising if they could be transferred directly to such use. In many cases the criteria used involve landform features that can not be resolved on Landsat images, or distinguished one from another.

There are three important aspects of satellite imagery for geomorphic investigation (Loffler and Sullivan 1979). First, the synoptic view provided on Landsat images is superior in most respects to aerial photomosaics. Second, the repetitive nature of the imagery allows seasonal and ephemeral changes to be observed. Third, the multispectral capabilities of the Landsat MSS allows recognition of features not seen on photos or by eye. With visual interpretation, Lofflet and Sullivan (1979 were successful in using these aspects of Landsat imagery to suggest that Lake Dieri, a former much enlarged lake, almost certainly occupied the basins that now contain Lake Eyre and Lake Frome.

Earlier, Loffler (1977) used visual interpretation of Landsat images to locate different landform types in the highlands of Papua New Guinea, and noted that bands 6 and 7 were best suited to geomorphic mapping.

Cochrane and Browne (1981) present a methodology for the systematic study of Landsat images to obtain geomorphic data. These workers, using Landsat-3 Return Beam Vidicon (RBV) images of the North Island of New Zealand, found a four stage approach successful. These stages were:

1. Photo analysis of tone, texture and lineaments on the RBV Images.
2. Comparison of these RBV data with topographic and geologic maps.
3. Photogeological mapping from conventional aerial photographs.
4. Field analysis of selected sites.

This approach allowed Cochrane and Browne (1981) to recognise three broad tectono-physiographic regions in eastern North Island. In addition, they were able to successfully map ten geomorphic categories that were easily recognisable on the RBV images.

One notable feature of all the work reviewed here is that Landsat imagery is used mainly to confirm prior landform distribution and classification. Even the more detailed work of Cochrane and Browne (1981) does not present a systematic landform classification. They map landform types that can be recognised on the imagery, without placing them in an integrated classification.

A CLASSIFICATION FOR USE WITH LANDSAT IMAGES

To be usuable with Landsat imagery, a landform classification must use criteria that are readily identifiable on Landsat images. As with aerial photographs, on Landsat images these are tone, texture, pattern and shape. Not all these properties are of value in the recognitiion of landform features on Landsat images. Some patterns, particularly lines, are of value, as are texture and shape (Table 1). However, tone is of little direct value in geomorphic interpretation as it is usually indicative of vegetation, not landforms.

Three kinds of lines observable on Landsat images are used in geomorphic interpretation. There are structural lineaments ridges, and rivers. Structural lineaments are easily seen on Landsat images and are a useful means of differentiating landforms with strong structural control from those with no structural control. However, care must be taken to remain objective. Because structural lineaments are almost always revealed in the landscape by erosional processes, it is reasonable to place areas with obvious lineaments in a major category, erosional landforms (Tables 1 and 2).

Ridges and rivers, or valley bottoms, can be taken together as differentiating criteria. On Landsat imagery there appear to be four major categories (Table 1). These are distinct ridges and rivers, rivers without ridges, ridges without rivers, and neither ridges nor rivers. This turns out to be a simplification of Ollier's (1967) classification of landscape profiles based on interfluve and valley profile shape (Figure 1). Further work will determine whether all of Ollier's classes can be distinguished on Landsat images. Table 1 suggests equivalents for different combinations of ridges and rivers.

A major problem with Landsat images is the difficulty in distinguishing between flat and gently rounded terrain. An associated problem is the recognition of drainage channels that are not large enough to affect reflectance values. Under these circumstances the location of ridges and valley bottoms may be difficult. Moreover, because of the lack of elevation data, it is difficult to distinguish between low plains and higher plateaux.

Texture can be used to distinguish between areas that have different densities of dissection (drainage densities), or repetative landform features. A simple breakdown between smooth, fine and coarse textures is suggested in Table 1, but particular Landsat scenes may allow further subdivisions to be made.

Areas with smooth textures can reasonably be assigned to landform classes with minimal relief differentiation, while other texture categories are assigned to appropriate landform classes. Very little work has been done on the limits which define different drainage densities, and the subject is not persued here.

Shape is important mainly because of the direct evidence it gives for specific landform types. It provides clues that can be used in association with other features in a landform classification.

In addition to presenting image features and their equivalent landsurface features, Table 1 also provides some geomorphic interpretation of the features. These interpretations were used in the development of the landform classification presented in Table 2. This classification is not intended to be exhaustive, but rather provide an easily adapted basis for further development. The classification is secondary to the main point of this paper, which is that remotely sensed images are a valuable tool in geomorphic interpretation.

Maps of landforms produced from Landsat images compare favourably with maps of a similar scale that could be produced using topographic maps, aerial photographs, or fieldwork. Fieldwork would be laborious and slow, and would not provide an overview. A similar comment can be made about aerial photographs which, although providing considerable detail, do not allow rapid broad scale mapping. Of the topographic maps available in Australia, only the 1:000 000 series gives equivalent details to that available from Landsat scenes. The 1:250 000 maps do not show as much fine detail, and the 1:25 000 maps are too detailed for regional mapping.

A major advantage of Landsat over other data sources thus lies in its synoptic view over a large area, but with enough detail to make accurate small scale landform mapping an easily accomplished task.

The above ideas were all developed using hard copy products from the Australian Landsat Station (ALS). However, experience with image analysis systems using computer-compatable tapes shows that enhancement techniques, especially filtering, can produce images which are superior to the hard copies available from ALS, and allow more precise mapping.

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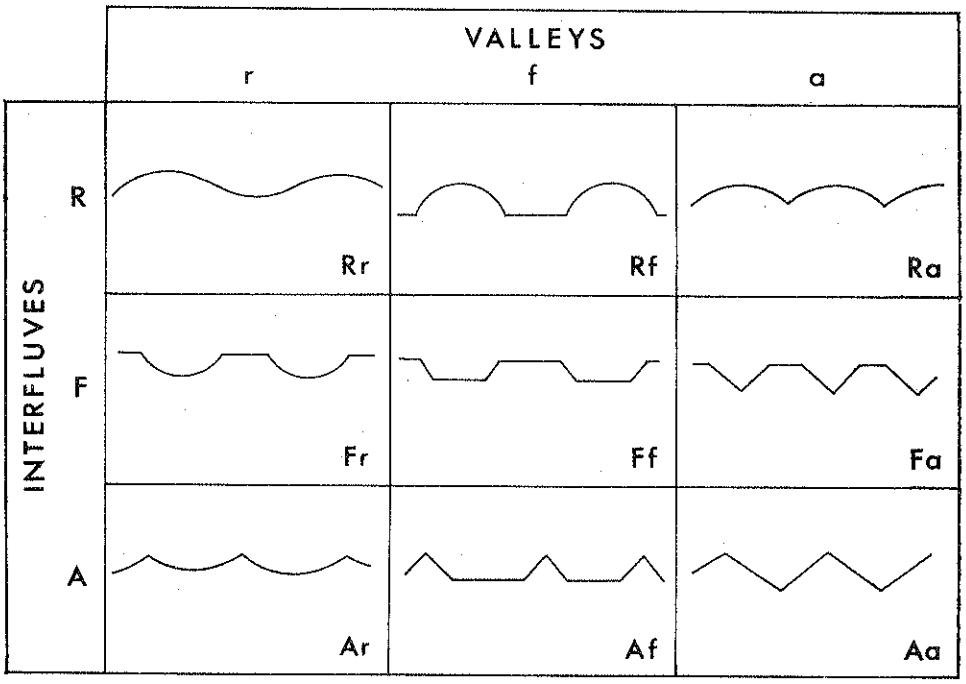

Figure 1. Classification of landscape profiles based on interfluve and valley profiles (from Ollier, 1967).

Table 1. Features observable on Landsat images, and their geomorphic interpretation

Image features	Landsurface features	Geomorphic interpretation
Lines	Structural lineaments	Erosional forms with structural control
	Ridges and rivers (Aa)*	Ridges and V valley forms
	Rivers without ridges (Ra, Fa)*	Dissected surfaces
	Ridges without rivers (Ar, Af)*	Surfaces with remnant upstanding ridges
	No distinct ridges or rivers (Rr)*	Low relief landforms with rounded forms
Texture	Smooth - very fine	Low relief, rounded forms
	fine	varying degrees of dissection and repetative landform features
	coarse	
Shape	Various	Usually contextural

* Symbols from Ollier (1967). See Figure 1.

Table 2. Landform classification for use with Landsat imagery

EROSIONAL LANDFORMS

- with structural control
  - Ridge and Y valley forms (sAa)
  - Dissected surfaces (sRa, sFa)
  - Surface with remnant ridges (sAr, sAf)
  - Plateaux and plainlands (sRr)
- without structural control
  - Ridge and V valley forms (Aa)
  - Dissected surfaces (Ra, Fa)
  - Surfaces with remnant ridges (Ar, Af)
  - Plateaux and plainlands (Rr)

DEPOSITIONAL LANDFORMS

- inland
  - Alluvial plains (Da)
  - Fans (Df)
- coastal
  - Coastal lowlands (Dc)

Note: Letters in brackets are suggested mapping symbols, taken in part from Ollier (1967) see (Table 1). An s prefix is used to denote landform with structural control, and a c or an f after a diagonal stroke refer to coarse and fine textures.

UPLIFT AND EROSION ON THE WEST COAST OF THE SOUTH ISLAND, NEW ZEALAND

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Marine and glacial outwash surfaces in the area west of the Alpine Fault offer surfaces whose ages are in some cases relatively closely dated, and others whose ages may be assumed within certain well-defined limits. Valley development below these surfaces can therefore be attributed to specific periods, and rates of downcutting quantified. In turn, amounts of erosion accomplished in the intervals between the development of successive terraces may offer an insight into rates and periods of uplift.

The paper examines these hypotheses in North Westland, in the coastal area between the Grey and Hokitika rivers, where a suite of Otiran outwash surfaces and late Pleistocene marine terraces and shorelines have been identified (Suggate, 1965).

Suggate, R.P. (1965), Late Pleistocene geology of the northern part of the South Island. N.Z. Geological Survey, bull. n.s. 77, Govt Printer, Wellington.

Taramakau R.

18 ka to present. 14-15 ka. 13 ka - present

3518 B. 8 380 B 2513 B

(linear erosion in river channel)

Karamea Riv. (small catchments)

120 ka 167 B

ANOMALOUS RED QUARTZ SANDS IN CERTAIN NULLARBOR PLAIN CAVES AND DOLINES

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SUMMARY

A preliminary account is given of the nature and occurrence of aeolian, red quartz sands from the eastern end of the Nullarbor Plain. Possible sources for these sands are discussed as are the implications of their restricted distribution and their possible palaeoclimatic significance.

INTRODUCTION

Speleological expeditions to the Nullarbor always return with more questions than answers; sand collected by Davey and Spate in 1982 in the course of other investigations has proved no exception. The small sample was examined by Jennings and Gillieson confirming a supposed aeolian origin. In November-December 1983 Gillieson, Jennings and Spate, assisted by J. Stockton, returned to the Plain to elucidate further the nature and distribution of the red sands. This expedition attempted to survey likely source areas of red sands; the stratigraphy and field characteristics were also investigated. A wide range of samples was collected for future study. Laboratory investigations are as yet incomplete but the unusual location of these sands and the bearing that this occurrence may have on arid zone palaeoclimate are thought to justify a preliminary paper here.

DISTRIBUTION OF THE RED SANDS

The first sample was collected from N99, a cave with two vertical "blowhole" type entrances, at the eastern end of the Plain. Collection was an after-thought to earlier observations in the nearby N98 and N97 caves where Davey and Spate had freely hypothesised a desert origin from the red colouration and from the mode of deposition. The 1983 expedition examined closely these and other caves and dolines as well as sand deposits widely in the Nullarbor Plain. Red quartz sands were found to be restricted to a small area toward the eastern end of the Plain; the five sites are all caves or dolines which form a compact group at the most ten kilometres apart, the Diprose Caves area (Fig.1).

During the 1983 trip we did not visit more northerly and westerly lying dolines and caves such as Old Homestead Cave (N83), Lynch Cave (N60) or Haig Cave (N55), but we did examine caves and sand deposits in all points of the compass from the Diprose area. We cannot be certain that red sands do not occur to the far WNW but following the mapping of Lowry (1970) and our field work we are confident that red quartz sands are absent from the coastal belt between White Wells (SA) and Cocklebiddy (WA). Once one is attuned to the red sands they are readily distinguishable from the usual buff sandy silts characteristic of the plains and cave deposits.

NATURE OF THE SAND

So far only the 1982 sample has been examined in the laboratory.

Particle size analysis (at 1 Ø intervals only) showed it to be well sorted with a strong mode in the fine sand class. These are aeolian sand characters.

Optical microscope examination showed that the sediment is dominantly made up of subrounded to rounded quartz grains, with a tendency to sphericity and much frosting of grain surfaces. The reddish colour derives from iron oxide held in small recesses on the surface of the grains. There is a secondary component of coarse calcareous grains causing negative skewness; they are dominantly angular to subangular. There is also a small component of heavy, opaque minerals, many of which are well rounded. This examination confirms an aeolian origin for most of the sample, though there is also a small clastic fraction from weathering of limestone, probably from the cave roof above this sample.

Scanning electron microscope examination of the acid insoluble fraction of the medium and fine sand classes confirmed the low surface relief of the grains attributable to prolonged abrasion during aeolian transport. Parallel ridges, especially along grain edges, are interpreted as remains of upturned plates, which are the results of collisions between saltating quartz grains during wind transport. These aeolian erosion effects are masked by secondary accretion of silica over grain surfaces in the form of amorphous pellicles and globules. Such coatings are known in dune sands elsewhere, e.g. the Simpson Desert, and are considered the result of pedogenic processes.

MANNER OF OCCURRENCE

Given the nature of the sand and the general absence of sand on the riverless surface of the Nullarbor Plain around the sites where it is found, there can be no doubt that it was brought to its present position by saltation in the wind. This does not exclude final emplacement by water because occasional rainstorms deluge into caves at high velocities.

Nevertheless, high angle planar beds seen in sections cut into the sands in N98 and N99 caves and the gross disposition of the sands as remnants of steeply sloping conical piles beneath entrances are incompatible with deposition from water falling down shafts from the surface but are consistent with air fall as dry sand. There is a sharp contrast between these red sand remnants and the low angle fans of lighter coloured sand being formed from the piles under the action of water coming down the shafts. These facts demonstrate that the red sands were emplaced by dry, single-grain fall directly under gravity. They were saltated to the holes in the cave roofs and fell into these traps.

In the N96 doline the sand has the form of a vegetated smooth blanket covering nearly the whole floor; exceptions are provided by the tops of blocks of limestone which have fallen in from the doline walls. The overall form of the sand surface is a gentle rise from the northern corner of the lozenge shaped doline to its southern corner. Augering and pits suggest that the rock floor of the doline rises from the northwest side to the SE side, where it is a shallow depression indeed. The blanket is thicker over this higher side of the floor. Over most of the floor the red sand sits abruptly on the former soil, a dark calcareous loam. The red sand has scarcely developed pedogenically, although there is some

darkening of the top few centimetres.

In all sites there are interbedded coarser sand laminae. However, there is a lack of evidence such as disconformities and soil horizons which might indicate substantial intervals in deposition.

The stratigraphy therefore points to the sand being deposited as a single and sharply defined event of short duration. The total amount of sand emplaced appears small after allowing for erosion of the cones in N98 and N99. Even in the case of N96, though the supply covered virtually the whole doline floor and concealed its irregularity, there was not enough to introduce more than a little into the cave from the doline. Thus the depositional event appears to have been limited in both the amount of sand transported and its duration.

DISCUSSION OF SOURCES OF ORIGIN

There are possible sources of sand in more or less all directions from the Diprose area. A preliminary examination of wind records from Cook (about 75 km NW), Eucla (about 200 km NSW), and Ceduna (about 250 km ESE), the nearest stations for which wind records are available, show that the sand-shifting wind resultant when calculated will be from a southerly quarter in all three cases with a westerly component in the case of Ceduna but possibly with an easterly component for the other two. At Cook, however, above-threshold winds from the northwest quadrant dominate in July (and possibly in other months of the winter quarter).

The possible sources of sand can be considered under three general headings:

(a) Residual soil from the Nullarbor Plain. The Nyanga Plain, part of the western end of the Eucla Basin, carries a thick cover of residual soil consisting chiefly of clay and kankar. This soil is absent from the rest of the karst and Lowry (1970) argues that it has been removed from the Nullarbor Plain by wind erosion. Jessup (1961) considered that the fine fractions of soils in northeastern South Australia and neighbouring parts of Queensland and NSW originated as calcareous dusts transported from the Nullarbor Plain. A small quartz fraction, as a residual from the limestone, is expectable and it could have lagged behind as saltation load.

Nevertheless this is an unlikely source for the Diprose sands. Lowry places this widespread deflation of the Nullarbor early in the Pleistocene. If this is correct, it cannot have incorporated the emplacement of these sands because in N96 they would have developed a more mature soil profile by this time. If the deflation were later, there are other strong arguments against such a source in that so much material is involved in this phase of deflation. Nearly the whole Plain was eroded to a depth of several metres and many more caves should have sampled the saltating load and been filled with aeolian material. Only a few caves have been affected and only small amounts of sand were trapped.

(b) Coastal dune sands. There is a discontinuous belt of coastal dune sands of various ages along the Nullarbor coast. In this study samples were collected as far west as the Roe Plains south of Madura where axial trends of parabolic dunes and major lobes point

to the Diprose area. This was done to obtain good samples of old coastal dune materials but not with the thought that materials from so far west had reached the Diprose area. Other coastal dune areas such as in the vicinity of Eucla and the Merdayerrah Sandpatch lie between these dunes and the Diprose Caves.

Closer to the Diprose Caves, at a distance of only 20 km, is the northern margin of the great mass of dunes which have advanced inland from the Head of the Bight. The axial trends are mainly from WSW to ENE as in the Roe Plains but near the Head of the Bight some advanced from SW to NE. These trends would carry sand past Diprose on its south.

There are smaller fringes of dunes along the top of the Bunda Cliffs between the Head of the Bight and the Merdayerrah Sandpatch. These could have fed the Diprose area following the WSW-ENE trend characteristic of the coastal dunes east and west of them.

Active dunes and younger, but fixed dunes, along the whole coastal belt discussed above contain a great, usually dominant, calcareous biogenic sand fraction. It does not seem possible therefore that any of these dunes could have supplied the Diprose area as the overall distances are not such as to allow the virtual loss of the calcareous fraction by differential rates of attrition, despite the less resistant nature of the calcite and aragonite dune sand.

Coastal dunes are, however, subject to leaching and older bodies of this nature are transformed into predominantly quartz remnants, though usually retaining layers of calcrete. Remobilisation of the "roots" of leached coastal dunes could segregate the quartz in a time of enhanced wind stress to produce a sand similar in mineral composition to the Diprose sands. Accounting for these sands by such a mechanism encounters two substantial difficulties. Firstly, how could the red colour be acquired in the short time suggested by their manner of occurrence? Reddening of quartz sand is usually considered a slow process (cf. Bowler and Magee 1978). The layers of quartz sand in the residual coastal dunes tend to be light pink, far removed in colour from the red sands. The second objection is that if the older sand bodies were remobilised, the neighbouring younger dunes of calcareous sand would certainly have been set in motion also. The lack of such sands at the Diprose Caves could be explained by postulating that remobilisation followed the leaching of the earliest coastal dune invasion so that its residuals only were available for fresh movement. This assumption again comes up against the lack of soil development in the N96 doline sand and thus its comparative youthfulness.

(c) Desert dunes of the Great Victoria Desert and its eastward continuation. Red quartz dunes lie in a wide arc north and east of the Diprose area - about 200 km in a NW direction, about 100 km to the N, NE and ENE. The red colour, dominant quartz composition and silica overgrowth of the Diprose sand so far examined invoke comparison with these desert sands. However, the axial trends of the dominantly linear dunes in this arc and the westerly aspect of the Y-junctions they make with one another show that these dunes, now largely fixed by vegetation, were emplaced by winds which would not have brought sand to the Diprose area.

If laboratory study of a range of sand samples sustains the field similarity between the Diprose and the desert sands and their unlikeness to the coastal dune sands, it must be proposed that the interior desert provided the materials. Despite comparative lack of knowledge of the inner part of the Nullarbor Plain it is apparent that few caves exist there but the possibility remains that more significant sand sites await discovery.

AGE AND SIGNIFICANCE OF THE RED SANDS

Preservation of such small bodies of dune sand in these karst niches so far from the source - if the desert origin is proved correct - has systematic interest in the field of desert geomorphology since some maintain there is only short range shunting action over the great lengths of desert dunefields. Wasson (pers.comm.) suggests that, rather than movements of dunes proper, 'sand streaks' (McKee 1979) may be involved in the Diprose occurrences.

This instance also possesses palaeoclimatic interest as the proposed source implies a different wind regime from that prevailing at the present time with a stronger northerly component. The last widespread phase of dune mobility was that between 25,000 and 13,000 BP (Bowler 1975); Bowler argues for a stronger northerly component in the wind regime of southeastern Australia at that time. Inactive linear dunes on Kangaroo Island continue the trend of the NW-SE quartz sand dunefield of the inner part of the Eyre Peninsula, itself an extension of the Great Victoria Desert. The present resultant of sand-shifting winds on Kangaroo Island is from the SW so these fixed dunes on the island imply the same sort of wind shift as has been suggested for the Nullarbor Plain above.

However, Wasson (1976) argues for a similar wind change to account for the trend of the Belarabon dunefield, which belongs to the Late Holocene. The lack of soil development in the N96 doline sands denies any great age for them, though pedogenesis must have proceeded slowly there. The rainfall cannot be very different from the 153 mm annual figure for Cook, the nearest station.

Dating the Diprose sands is therefore of some moment; samples collected for thermoluminescence dating are now being processed in the Department of Physics, ANU.

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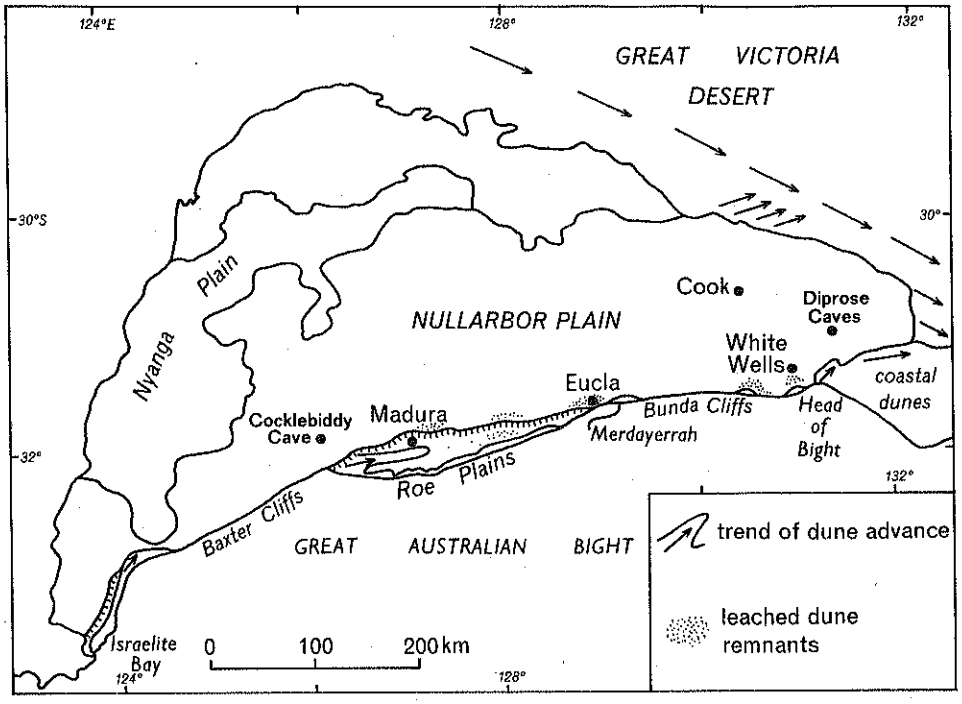

Figure 1 Map of Nullarbor Plain, showing location of places mentioned in text.

THE LONG SLEEP OF HISTORICAL GEOMORPHOLOGY

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Australian and New Zealand Geomorphology Group
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ABSTRACT

Three parallel sub-disciplines contribute to the field of earth history: historical geology, historical pedology and historical geomorphology. In historical geology a coherent code of principles called the stratigraphic code was set up more than forty years ago. These principles are relevant to the other sub-disciplines, although the word stratigraphy is misleading.

In historical geology the basic mapping unit is the formation; in historical pedology the pedoderm. There is no such unit in historical geomorphology. Its absence prevents proper critical review of field work and correlation. Serious debate is long overdue.

The terms 'landform association' and 'geomorphic surface' have previously been proposed. A unit erected for the purpose of throwing light on time relationships of past events must not be defined in terms of those time relationships, but on observable physical features.

INTRODUCTION

Few geomorphologists doubt that their discipline contributes to the understanding of earth history. What, then, are the geomorphological units or entities that can be put forward as evidence of events in earth history? Historical geologists discuss formations; since 1970, historical pedologists can discuss pedoderms. What do historical geomorphologists discuss? No suitable term is in use.

STRATIGRAPHIC PRACTICE AS A MODEL

In studies of Pleistocene glaciation, for example, geomorphological evidence is marshalled under the name of a glacial Advance. This is an inferred event, just as a marine transgression is an inferred event. By contrast, a geologist would postulate a marine transgression on the evidence of a named formation that he had first defined, described, and mapped. This procedure, formalized in stratigraphic codes, ensures that the evidence for historical inferences is open to criticism and re-evaluation.

Formations, together with groups, members, etc., are the rock-stratigraphic, or lithogenetic units that the geologist finds in the field. The Code recommended by the American Commission on Stratigraphic Nomenclature (1961), which was based on the proposals of Schenck and Muller (1941) 20 years earlier, clearly separated such rock-stratigraphic units from time-stratigraphic units (systems, series, stages) which are bodies of rock defined to be the record of a specific interval of geologic time. Both units were distinguished from geologic-time units (eons, eras, periods, epochs, ages). This three-fold scheme of units is a key feature for the orderly discussion of earth history.

A fundamental concept expressed in the Code of 1961 is that formations, defined for the purpose of elucidating time relationships must not be defined in terms of time relationships. Formations are distinguished, delimited, recognized and defined by observable physical features: lithological texture, structure, and chemical composition. Inferences concerning time, both time-sequence and contemporaneity, although primary objectives of stratigraphic studies,

are specifically excluded from a definitive role. The fact that a formation may have been deposited at different times at different places is not the main point. It is simply improper to define a unit from which an event is to be inferred in terms of the occurrence of that event.

SOIL STRATIGRAPHY

Soils as materials for the study of earth history have been persistently misunderstood by geologists. The rigorous logic that had been applied to the definition of formations was not applied to so-called 'soil-stratigraphic units'. An Australian proposal (Brewer, Crook and Speight 1970) was the first attempt to adapt Schenck and Muller's concept of a time-significant objective unit to soil-historical studies. This resulted in the definition of a pedoderm. Whereas a formation is a body of rock, a pedoderm is a mantle of soil. In each case the scientist seeks and documents evidence of recognizability, homogeneity, continuity and boundary conditions that will eventually contribute to the inference of an event in earth history. Wishful thinking is checked by the requirement that the designation of the unit is to be published and the physical evidence will be open to any amount of future research.

ASPECTS OF EARTH HISTORY

It is unfortunate that the term 'stratigraphy' has been inappropriately extended to 'soil-stratigraphy'. Stratigraphy deals with layers of material placed one upon another. Pedoderms are not placed on each other in this manner. They are 'ground-skins'. The skin is developed at the surface of the ground and extends some distance below the surface to a diffuse lower boundary. The need to separate the concepts of pedoderm and stratigraphic formation has recently been re-stated by Beckmann (1984) and by Walker, Beckmann and Brewer (in press) in opposing a re-definition of 'pedoderm' by Butler (1982). The study of pedoderms is better called historical pedology, just as stratigraphy, together with paleontology, falls within the field of historical geology.

To extend the discussion to include geomorphology, it would be even less appropriate to speak of landform stratigraphy than of soil stratigraphy. Although fundamental concepts relevant to historical studies of landforms and soils were first developed by stratigraphers these concepts are not specific to the study of strata, but apply more generally.

Historical geomorphology, historical pedology, and historical geology can be considered as parallel sub-fields, each yielding evidence on the nature and sequence of events in the remote past. They are independent sub-fields of earth history.

GEOMORPHOLOGY AS EARTH HISTORY

Past contributions of geomorphologists to the study of earth history were often termed 'denudation chronology'. From the theoretical cycles of landscape development of Davis and Penck, it was deduced that relict, generally gently sloping surfaces and landforms might be identified in the landscape as having escaped destruction by the later erosion cycles. Numerous such relict surfaces were recognized and it was the practice to give names to 'peneplains' reconstructed by linking correlated remnants. It seems that elaborate time-sequences and wide-ranging correlations were sometimes erected on too little verifiable evidence. At all events geomorphologists did not establish formal principles to ensure that verifiable evidence was documented as the stratigraphers did.

The need for such principles was stated by Leland Horberg in 1952 in words that are an indictment of the succeeding 30 years of stagnation in geomorphological thinking. He said that the historical approach in geomorphology

> "imposes the responsibility of developing a body of principles by which geomorphic history may be interpreted. These principles as such are seldom considered and are yet to be defined and organized.... It may also be desirable, in search of new approaches, to apply familiar stratigraphic concepts leading to such topics as: (1) the nature and significance of geomorphic unconformities; (2) criteria for determining relative age of

cyclical surfaces; (3) bases for correlation of erosional and depositional surfaces; and (4) the concept of facies applied to landscapes." (Horberg 1952).

The lack of perceptible progress along these lines cannot be due to ignorance of this passage, for it is quoted in full on p.32 of 'Principles of Geomorphology' by Thornbury (1954).

The landscapes in which historical geomorphology is most straight-forward, and can make its greatest contribution compared to historical geology and historical pedology are those of multiple glaciation in the Pleistocene. In my thesis study of such an area at Lake Pukaki, New Zealand (Speight 1963) I insisted on naming and formally describing the assemblages of landforms that I observed, rather than either the glacial advances that I inferred from them, or the poorly exposed and lithologically monotonous tills and gravels beneath the landforms. Having no precedent to follow, I set up a type of basic historical-geomorphic unit modelled on the formation concept and defined thus:

"A landform association is a subdivision of a landscape in which the landforms have such an ordered arrangement, consistency of slopes, uniformity of erosional development, and degree of obliteration of detail as would indicate that they originated together as a land surface." (Speight 1963, p.162).

There are many deficiencies in this definition due to a lack of discussion or debate at the time. To this day, no geomorphologists studying glacial history in New Zealand or elsewhere have acknowledged that geomorphological units should be defined to facilitate discussion of the validity of field work and correlation.

R.V. Ruhe carried out a number of studies in which 'geomorphic surfaces' were mapped and named. His attempt at formal definition, however, is full of problems. It demands a lot of accurate data, and it is more analogous to a time-stratigraphic unit than a rock-stratigraphic unit:

"If formality is desired, geomorphic surface may be used, and it is a portion of the land surface that is specifically defined in space and time. It may occupy an appreciable part of a landscape

and may include many landforms. It may also include many landscapes. The geomorphic surface must be a mappable feature. Its geographic limits and distribution in elevation must be delineable on aerial photographs or topographic maps. It has geometric dimensions which must be specified and which may be analyzed. It must be defined in association with other geomorphic surfaces in order to place it properly in its spatial and time sequence. It is associated with bedrock or sediments and may have bedrock or sediments associated with it. These associations must also be specifically defined in space and time. It is datable by relative or absolute means, and its dating must be specified. After definition of all of these relationships, the geomorphic surface is labeled, usually with a geographic name." (Ruhe 1969).

B.E. Butler's term groundsurface (Butler 1959) is a complex time-rock-soil unit that incorporates the concept of a geomorphic surface. Its use as a basic geomorphological field and mapping unit is vitiated both by the use of time as a definitive attribute (by way of the K-cycle) and by the incorporation of soil and sediment attributes. The term pedomorpholith proposed by van Dijk, Riddler and Rowe (1968) is similarly unsatisfactory in attempting to subsume three disparate kinds of evidence in a single unit. Such a unit invites sleight of hand.

SURFACE DEGRADATION

Except for some buried terrains, all of the features identified as former land surfaces occur at the present ground surface. They have thus been consistently exposed to geomorphological processes tending to destroy their original form. This makes landforms much more difficult to use as indicators of events than are sedimentary rocks. The moment of final burial of a rock formation leaves it protected from surface processes.

Strictly, the diastem, disconformity or unconformity forming the upper surface of any formation is a geomorphic surface. In this, geomorphic surfaces relate to diastems just as soils do (Morrison

1967). Since most diastems remain buried, they can be observed mainly in cross-section in favourable exposures, where the skills of the stratigrapher and the historical pedologist are more appropriate. The geomorphologist's forte is the mapping of surface features. The fact that no present surface feature is wholly relict must be faced, and set against the wonderful accessibility of landforms by comparison with the fragmentory observations of rocks and soils.

FEATURES AND PROCESSES

The features enabling one to recognize a geomorphic surface will generally relate to changes in the kind or severity of earth surface processes. They may reflect aridity versus humidity, glacial, periglacial or fluvial activity, erosion contrasted with deposition, subaerial versus marine action, high water table versus low water table. Each causative phenomenon calls for the identification of diagnostic features. They may appear in the drainage network, the hillslope profile, the characteristic suite of landform elements and other attributes of landform patterns.

COMPLICATIONS

The recognition of geomorphic surfaces remnant from a former landscape is complicated by the influence of rock structures. Young (1977, 1978) has shown how previous workers had misinterpreted structural features near Sydney as remnants of former land surfaces. That is not to say that there are no such remnants in that area. The landform development on the coastal and inland faces of the Budawang Range near Braidwood is so different that it is surely useful to hypothesise that the inland face is a remnant surface. The likelihood that this surface has lost several metres of material since the Eocene by an erosion process that is still active is scarcely sufficient to invalidate its status as a remnant surface. Parts of the adjacent coastal face have lost about 600 m in the same interval.

In this case, as in most problems of historical geomorphology in Australia, there are alternative hypotheses to

consider. This emphasises the need for orderly marshalling of the landform evidence from which earth history is to be inferred.

CONSTRUCTIVE DEBATE

Geomorphologists may have invited criticism in the past for their simplistic models of former landscapes. Coincidence of altitudes is scarcely enough in itself to build a peneplain on. However, the perceived continuity and functional relationships of parts of the terrain seen on an air photo or satellite image should not be dismissed as trivial. Such perception depends on geomorphologists' increasing knowledge about how landforms work. The reconstruction of past landscapes is a valid application of it.

It seems to me that the time is overdue for some serious debate about the materials of historical geomorphology. We infer epochs of aridity, of glaciation, of enhanced fluvial activity, and of extreme landscape stability associated with deep weathering. What about the material evidence on which the inferences are based? Should we not require of each other that such evidence should be documented to an agreed standard? A more rigorous approach to the presentation of landform evidence for historical studies, including the definition of a basic mapping unit, would be more than belated housekeeping. The body of principles that Horberg spoke of half a lifetime ago may begin to emerge, prompting a revitalisation of historical geomorphology.

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BIOPHYSICAL RESPONSES TO STREAMFLOW IN BLACKWATER CREEK, CENTRAL QUEENSLAND

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In October 1982 a year-long freshwater monitoring programme was commenced along Blackwater Creek (200km west of Rockhampton)(Figures 1 and 2). The study was undertaken to provide Curragh Queensland Mines Limited with baseline sediment, water quality and aquatic biological data, necessary for the company to assess potential impacts of the newly operating coal mine north of Blackwater township. Measurements were made of 30 water quality parameters, 13 sediment chemistry parameters, channel form and sediment characteristics, and benthic invertebrate populations (Table 1). Stream runoff data was obtained from the Queensland Water Resources Commission.

Blackwater Creek is an intermittent stream characterised by extremes from high flood runoff to isolated pools. Mean annual rainfall at Blackwater Township is 580mm. During the study period 76.5% of the runoff was concentrated in 12 days in April-May 1983 a maximum instantaneous flow of $203m^3s^{-1}$ and a maximum daily discharge of 21 800ML. (Figures 3 and 4). For much of the remaining time no flow occurred. These variations in the rate and volume of runoff place a number of constraints on the biogeochemical processes and ecology of the stream.

The ionic composition in waters of Blackwater Creek are variable and complex. Na^+ and Cl^- are dominant though the anion SO_4^{2-} is commonly co-dominant with Cl^-. Salinity measured as conductivity at most sites does not appear to be highly variable on a seasonal basis though the reference site does. (Figure 7). Alkaline conditions (pH 7.1-9.3) occur throughout the year.

The bed sediments of Blackwater Creek are predominantly coarse sands with little variation in particle size downstream. Blackwater Creek is, however, a pool/riffle stream where there is a tendency for slightly coarser sediments to accumulate at riffles (Figures 5 and 6).

The dominant influences on water quality are the intermittent (and) variable flow regime and persistent sewage waste discharges from Blackwater township. Consequently, a feature of the creek's water quality is extreme temporal and spatial variability, particularly in dissolved oxygen, macronutrients, faecal coliforms and algal biomass. Total P, $N\text{-}NH_4^+$ and $N\text{-}NO_3^- + NO_2^-$ concentrations

increase significantly at the sewage outflow then decrease downstream (Figure 8) while mean dissolved oxygen data reveal a classic 'oxygen sag'. There are also significant increases in Mn, Ni, Co, total P and Fe in the pool sediments at the sewage outflow.

Thirty-eight taxa of benthic invertebrates were identified, consisting mainly of chironomids and the oligochaete Branchiura sowerbyi. Stream pools below the sewage outfall usually contained the highest number of individuals and species. Benthic community characteristics are significantly affected by conductivity, dissolved oxygen, Mg, Fe and total P in water, and by Cu, Ni, B, Mn, Mo and total P in sediments (Table 2). Flooding in April-May 1983 led to a significant reduction in all community parameters except biomass, though recovery to pre-flood conditions occurred in less than four months.

TABLE 1

BASELINE MONITORING STUDY PARAMETERS, SITES AND SAMPLING FREQUENCY
BLACKWATER CREEK, CENTRAL QUEENSLAND

PARAMETERS	SITES	SAMPLING FREQUENCY
1. SOILS		
Soil Samples	12	Sampled once in October 1982
Soil texture	12	Samples 0-10cm and 40-50cm
Soil salinity	12	Composite samples 0-50cm
2. CHANNEL BED SEDIMENTS		
Sampling Sites	12	Sampled once in October 1982
Particle Size Distribution		
Pools	12	
Riffles	12	
3. TRACE ELEMENTS IN POOL SEDIMENTS		
Fe	7	4
Mn	7	4
Al	7	4
Cu	7	4
Zn	7	4
Ni	7	4
Cr	7	4
Mo	7	4
Co	7	4
Cd	7	4
B	7	4
P	7	4
As	7	2
Hg	7	2
Pb	7	2
4. WATER		
Physical		
Temperature	7	6
Turbidity	7	6
Non-filterable Residue (suspended solids)	7	6
Conductivity	7	6
Chemical		
Dissovled Oxygen	7	6
pH	7	6
Total Dissolved Salts	7	6

PARAMETERS	SITES	SAMPLING FREQUENCY
4. WATER (cont.)		
-Major Cations		
Na^+	7	6
K^+	7	6
Ca^+	7	6
Mg^{2+}	7	6
Major Anions		
CO_3^{2-}	7	6
HCO_3^-	7	6
Cl^-	7	6
SO_4^{2-}	7	6
Macronutrients		
Total Phosphorus	7	6
Nitrogen (NH_4^+)	7	6
Nitrogen ($NO_3^- + NO_2^-$)	7	6
Chlorophyll-a and Pheophytin-a	7	6
Trace Elements		
Fe	7	6
Mn	7	6
Al	7	6
Cu	7	6
Zn	7	6
Ni	7	6
Cr	7	6
Mo	7	6
Co	7	6
Cd	7	6
B	7	6
As	7	6
Hg	7	6
Pb	7	6
Microbiological		
Faecal Coliforms	7	6
5. BIOLOGICAL		
Benthic Invertebrates		
Taxonomic identification	7	3
Structure/Diversity Relationships	7	3

TABLE 2 : REGRESSION TESTS OF BIOLOGICAL INDICES ON SIGNIFICANT WATER QUALITY PARAMETERS FROM BOTH OCTOBER 1982 AND FEBRUARY 1983 SURVEYS

Dependent Variables-Biological Indices	Independent Variables - Significant Water Quality Parameters								
	Susp. Solids	DO % Sat	Na^+	Mg^{2+}	pH	Cond.	Fe	Total P	NH_4^+
H'		* (+)					** (-)	* (+)	
No. of Species				* (+)				** (+)	
Richness				* (+)				*** (+)	
Evenness-J							** (-)		
Biomass					* (+)			* (+)	

* p < 0.05 ** p < 0.01 *** p < 0.001

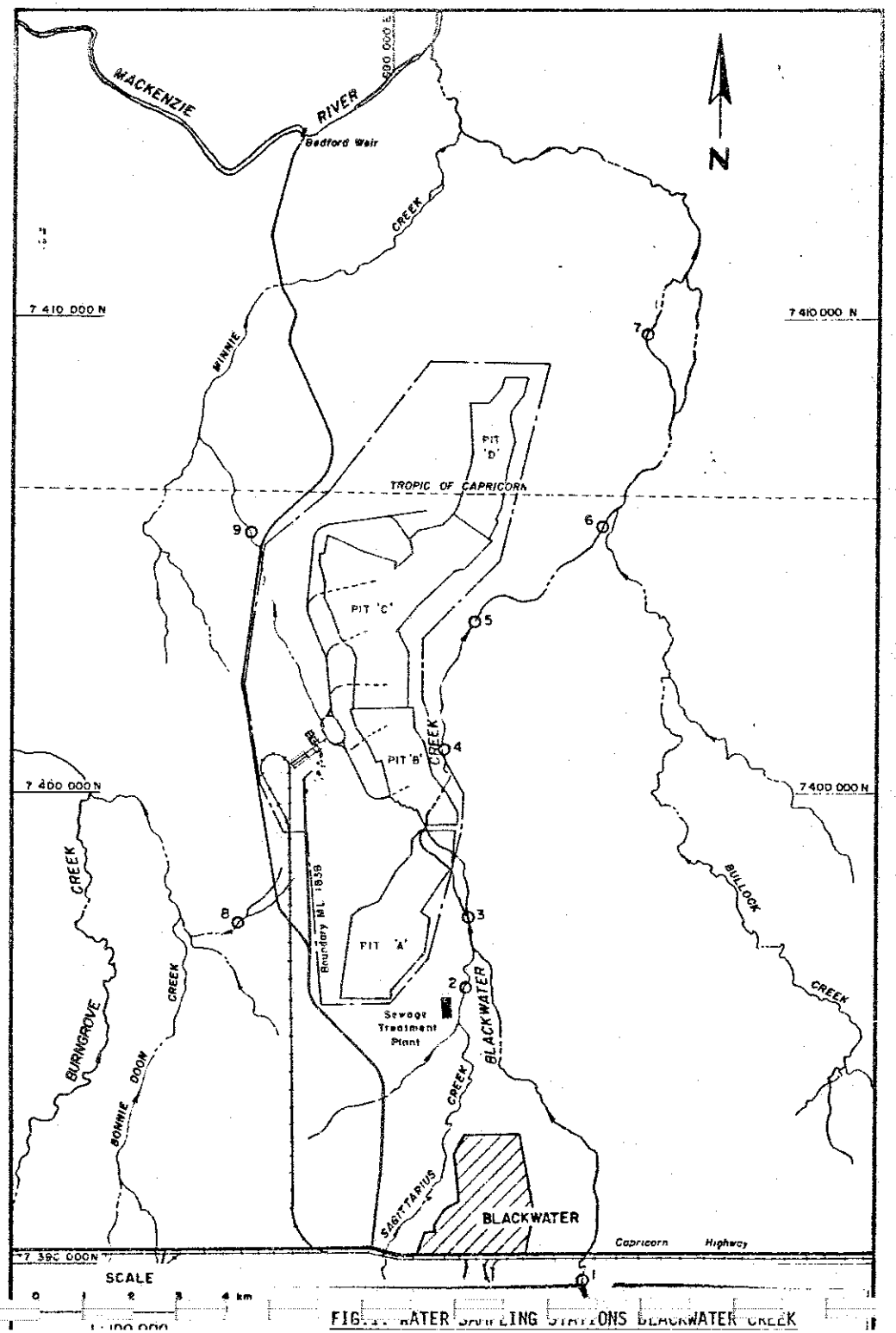

FIG. 1: WATER SAMPLING STATIONS BLACKWATER CREEK

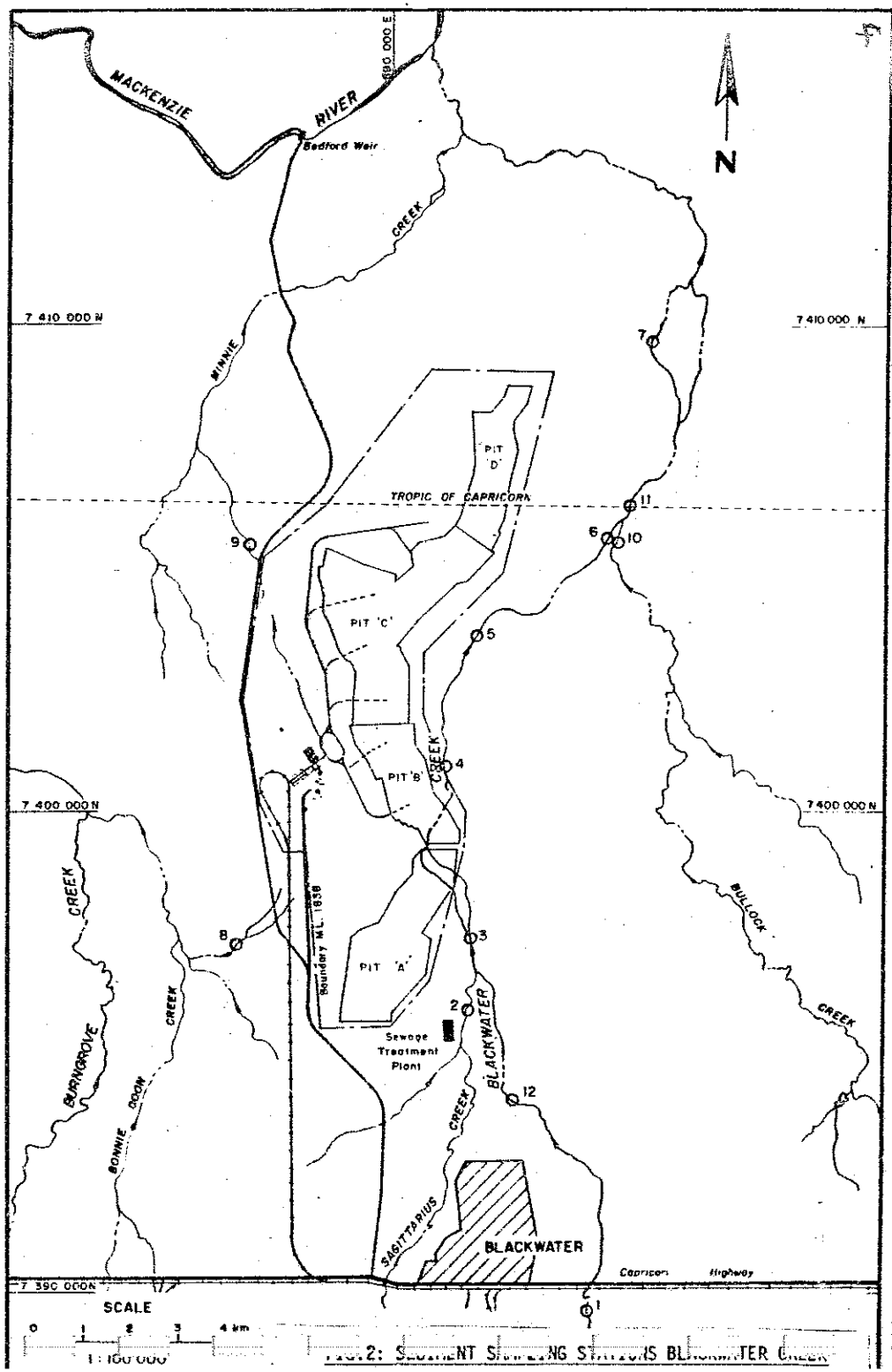

FIG. 2: SEDIMENT SAMPLING STATIONS BLACKWATER CREEK

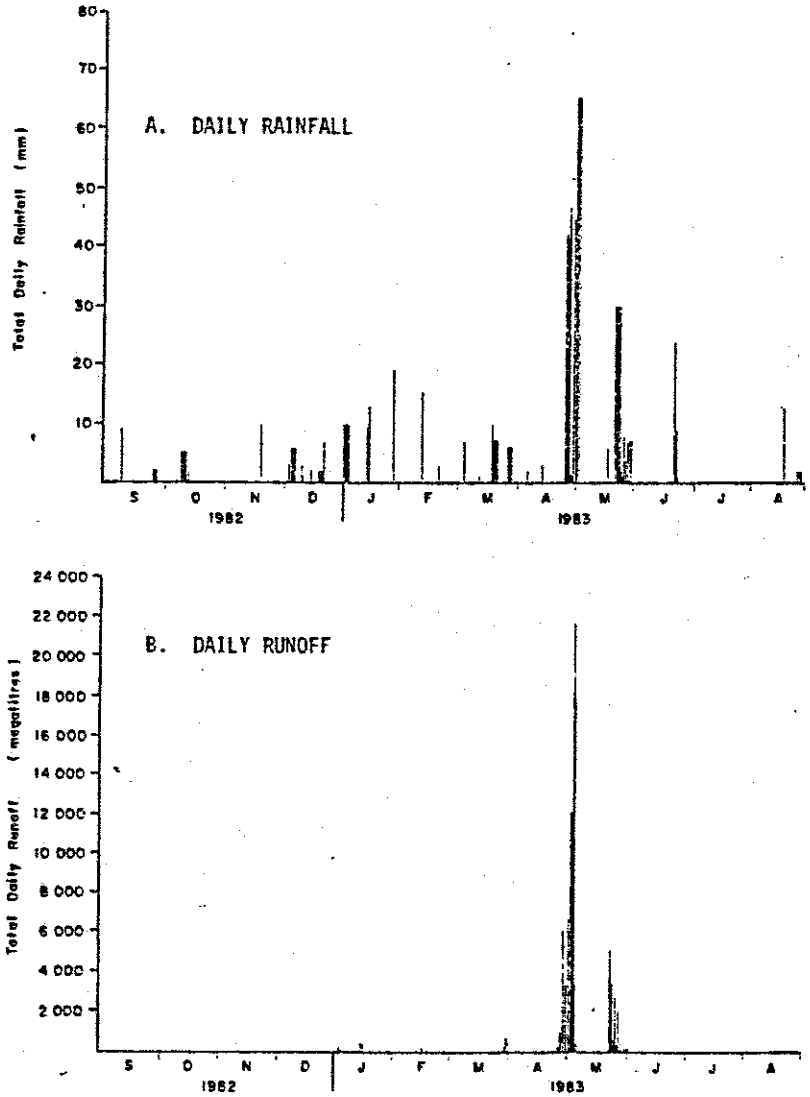

FIG.3: DAILY RAINFALL (BLACKWATER) AND BLACKWATER CREEK RUNOFF (CURRAGH), SEPTEMBER 1982 TO AUGUST 1983

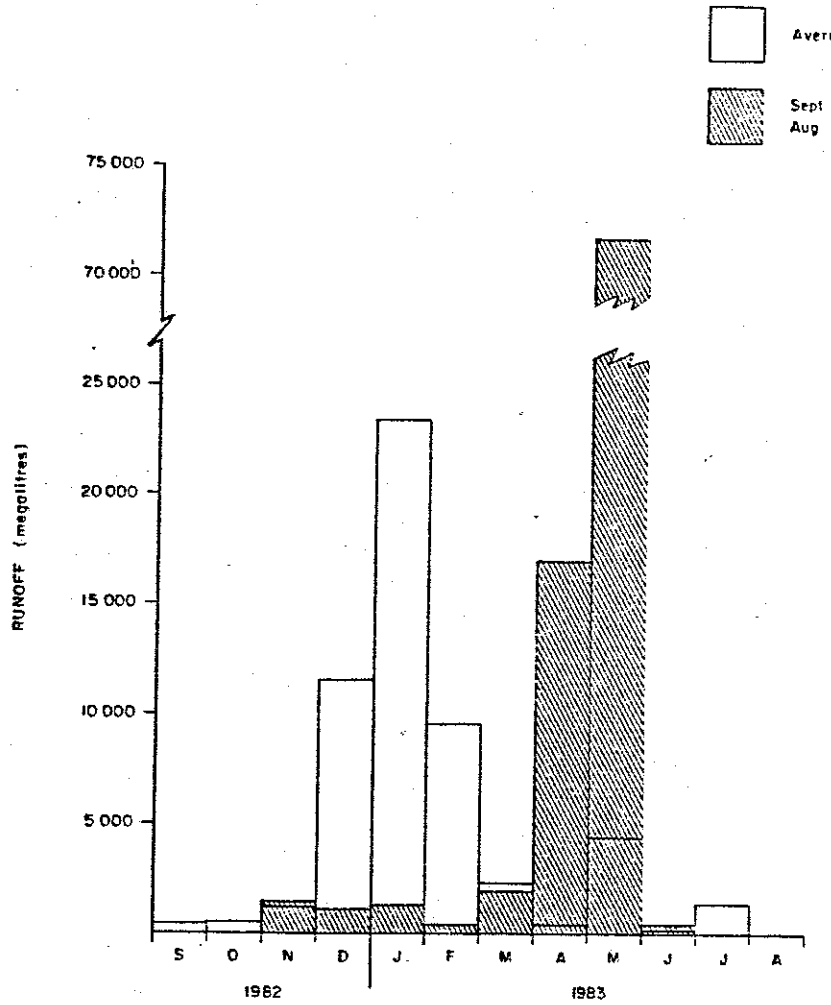

FIG.4: COMPARISON OF BLACKWATER CREEK AVERAGE MONTHLY RUNOFF WITH RUNOFF DURING SAMPLING PERIOD

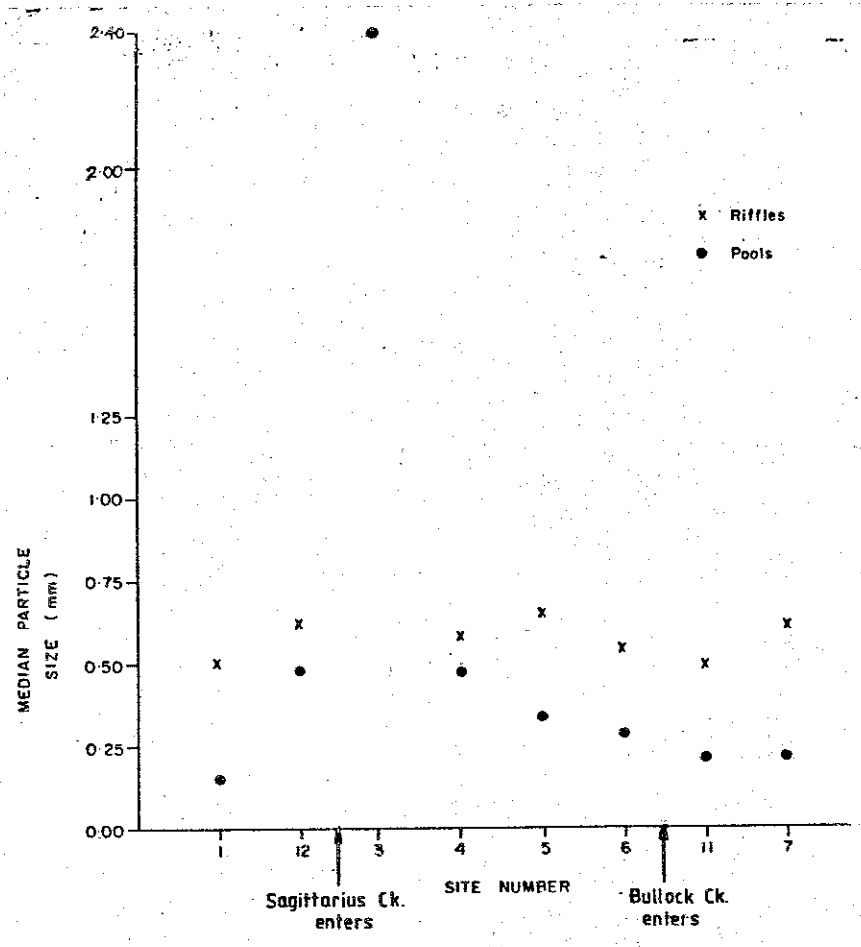

FIG.5: MEDIAN PARTICLE SIZE OF CHANNEL BED SEDIMENTS AT POOLS AND RIFFLES ALONG BLACKWATER CREEK

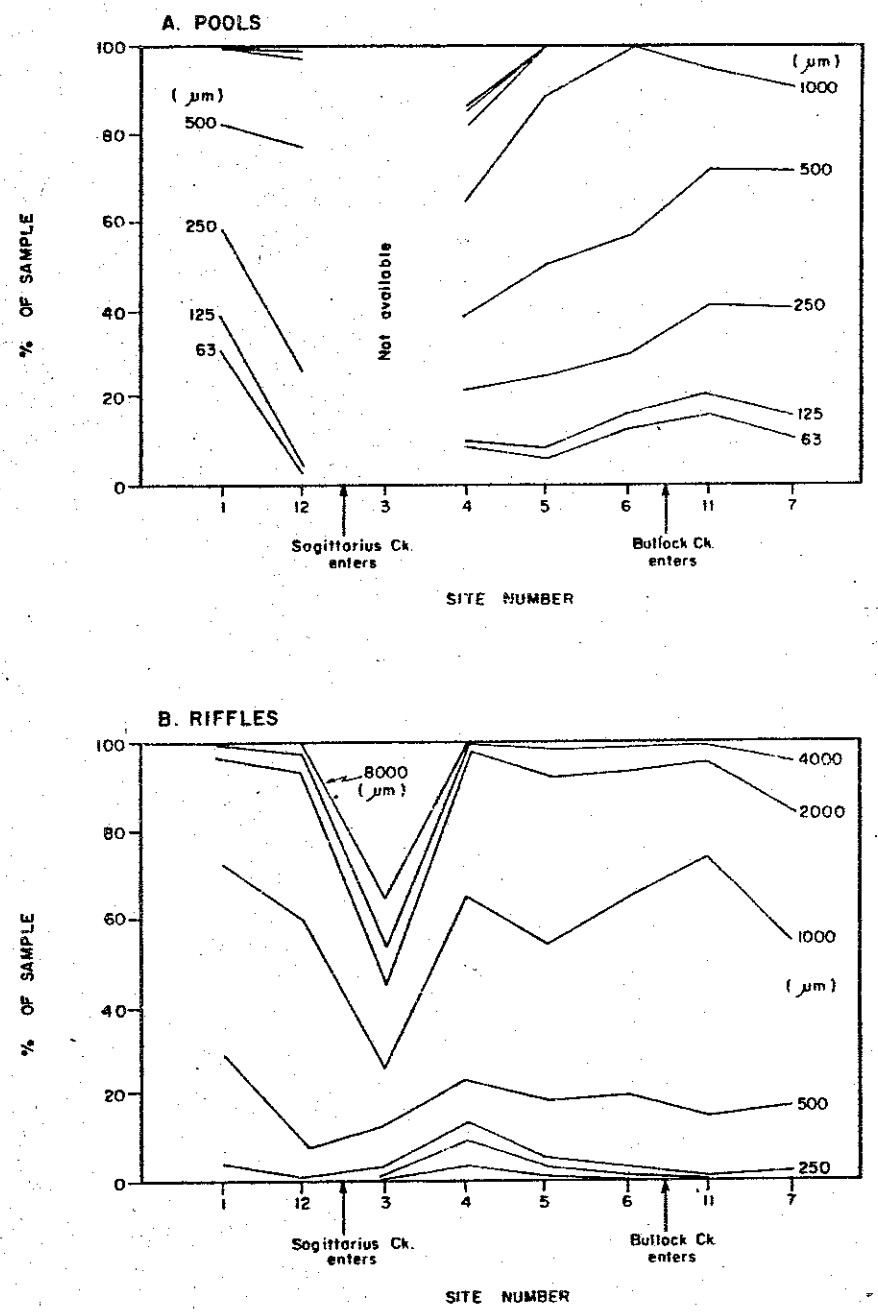

FIG.6: PARTICLE SIZE DISTRIBUTION OF CHANNEL BED SEDIMENTS ALONG BLACKWATER CREEK

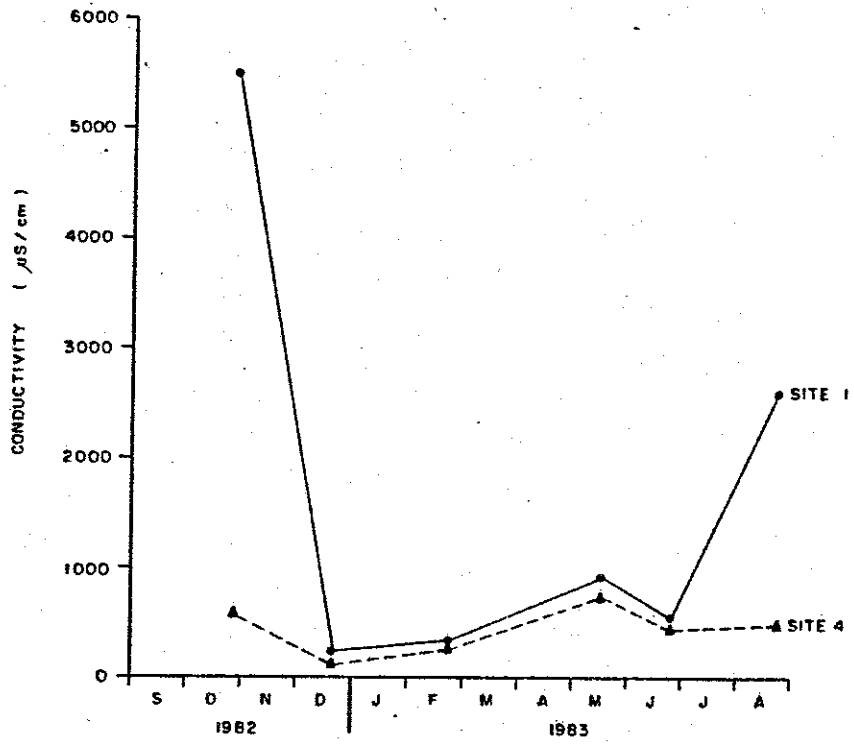

FIG.7: SEASONAL VARIATIONS IN CONDUCTIVITY IN BLACKWATER CREEK

-- SITES 1 AND 4 OCTOBER 1982 TO AUGUST 1983

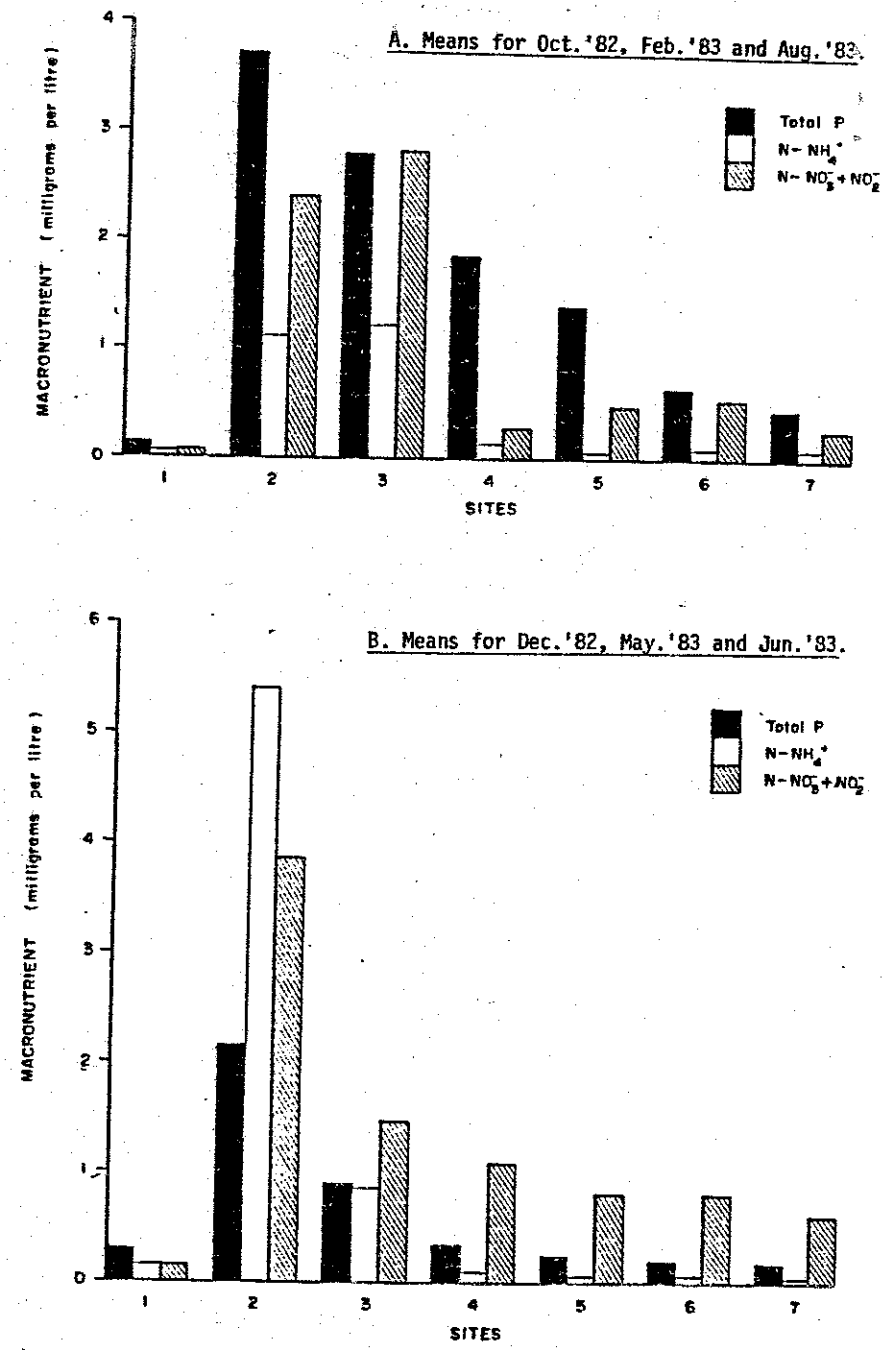

FIG.8: MEAN CONCENTRATIONS OF MACRONUTRIENTS IN BLACKWATER CREEK

RELATIONSHIPS BETWEEN SURFACE GRAVELS AND STAGES IN SOIL GEOMORPHIC PLANAR DEVELOPMENTS, TARA PLAINS, S.E. QLD

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Four different types of gravels in local accumulations at the surface and/or within the soils on the Tara brigalow plains (Fig. 1) show systematic distribution in relation to geomorphic subdivisions. These subdivisions (shown in Fig. 2) occurring in characteristic sequence in valley cross section (Fig. 3), have been related to particular stages in landscape development (van Dijk, 1984).

The following descriptions of the gravels also give brief site descriptions; type sites are indicated in relation to the geomorphic situation in Fig. 2.

1. Wangaby small pebble gravel. Dark red brown to purplish red brown, often highly glossy surface; irregular subangular, in size 0.5-5cm (dominating size 1-3cm); moderately hard, strongly ferruginous, undifferentiated internal fabric; fine irregular voids are common, often with dark purplish cutans.

Occurs on broad, flat or slightly indented interfluve ridges on pediment-like sloping plains of the landscape unit "miscellaneous high lands".

Type site: south-eastern part of parish Wangaby, e.g. quarry along Windermere Road at 256597, topographic map Surat 1:250,000, in a slight depression in a broad, flat topped interfluve (this interfluve with its hardened fringes is apparently the remnant of an ancient "duricrusted" drainage path, and represents a typical example of inversion of relief for this type of landscape - K.G. Grimes, Mines Dept, Brisbane pers. comm.).

2. Moonya medium to large pebble gravel. Pale grey with diffuse dark red brown flecks, rough surface, irregular subrounded, in size 1-12cm (dominating size 1.5-4cm); very hard with crystalline fracture faces, seemingly silicified with varying degree of ferruginization, generally undifferentiated internal fabric but often with a conspicuous 1-1.5mm thick dark red brown ferruginous rind.

Occurs in small areas of the "upper clay plains", type site at 304586 topographic map Dalby 1:250,000.

3. Mulleellee very small pebble gravel. Red brown to dark red brown smooth surfaces, subangular to subrounded, the smaller sizes subrounded to rounded, generally 0.1-3cm (dominating size 0.1-1cm); moderately hard with earthy, undifferentiated interior; smaller sizes are strongly magnetic.

Occurs in large amounts on very gently sloping, usually slightly depressed areas of sandstone land, generally associated with shallow red earths, situated on the flanks of landscape units of the "upper clay plains" and on the low broad interfluves of the "sloping plains".

Type site at 336597, topographic map Dalby 1:250,000.

4. Coowonga very small pebble gravel and coarse sand. Uncoated ("clean"), shiny, well-rounded, generally 0.2-0.5cm in size; usually quartz, and obviously derived from well-sorted alluvium. Conspicuous accumulations are associated with ant mounds on deep, heavy clays in some areas of the "lower clay plains", and are concentrated following dispersion of the clay material.

Type site at 342602, topographic map Dalby 1:250,000.

Gravel types 1, 2, and 3 seem to be of specific accretionary origins. Both the Wangaby and Mulleellee gravels are undoubtedly concretions showing well-developed concentric structures. The Moonya gravels are usually rather structureless and could be seen as alluvially rounded silcrete fragments. However, lack of erosional scratching and the irregularly rounded, often tubelike, forms support the interpretation of accretionary origin.

Gravel type 4, the Coowonga gravels, are sorted alluvial grains. This type attracts attention because the grains are clean due to the absence of cutans which are often present on alluvial gravels and sand.

The plains, each of different geomorphic character, on which the gravels occur, have been related to three successive stages of Tertiary (Mid Miocene to Mid-Late Pliocene) planar landscape development called Cobba-doomally, Erringibba, and Tullagrie (van Dijk loc. cit.).

Both the Wangaby and Moonya gravels seem to be associated with the Cobba-doomally stage which was characterized by major valley pediplanations resulting in large-scale regional valley plains leaving only scattered, isolated residual complexes ("miscellaneous high lands" in Fig. 2). The Wangaby gravels occur on pediment-like slope sections of the latter residuals graded to the Cobba-doomally valley plain level. The Moonya gravels are found on remnants of these valley plains ("upper clay plains" in Fig. 2).

The Mulleellee gravels occur on the planar land forms related to the Erringibba stage. This stage was characterized by the cutting of wide, though usually shallow, valley troughs which dissected and extensively cut away the Cobba-doomally surface.

The Coowonga gravels are typical for the alluvia of deep and extensive cut-and-fill formations related to the Tullagrie stage. These formations are often several metres thick with widths of several, up to 20 kilometres and developed as major modification of the previous Erringibba valley floors.

The significant differences in character and in geomorphic relationships between the three types of accretionary gravels are in strong contrast to the similarities of the sites of occurrence of the gravels with respect to local environmental conditions such as topography and bedrock. Apparently the determining factor in the cause of differences in gravel development were soil weathering regimes governed by varying climatic-environmental conditions prevailing during the respective soil geomorphic stages. Understanding of these past conditions can be acquired from available geochemical knowledge concerning the formation of gravels (see e.g. Schwertmann and Taylor 1977).

Finally, the Coowonga gravels are notable with respect to geochemistry because they indicate, though indirectly, unique soil weathering characteristics of the thick clay mantles which developed during the Tullagrie stage through two diagnostically important features: a) the grains which are clean mentioned above; this indicates soil conditions unfavourable for cutan development which is often common on alluvial gravel and sand; b) the difficulty in discerning the grains, even if present in large amounts, in field samples; this is apparently the result of particularly intensive lac plasma development in the soil profile which results in firm embedding of any coarse fractions in a very fine clay mass.

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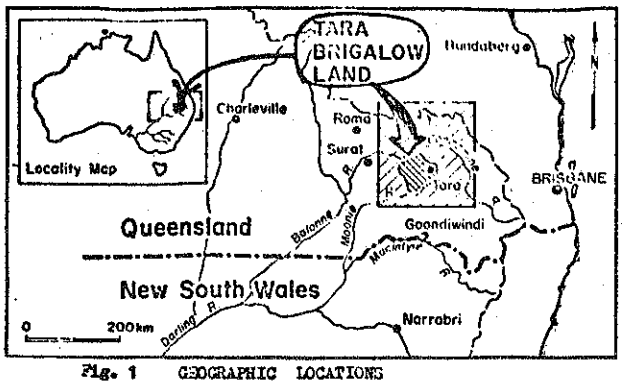

Fig. 1 GEOGRAPHIC LOCATIONS

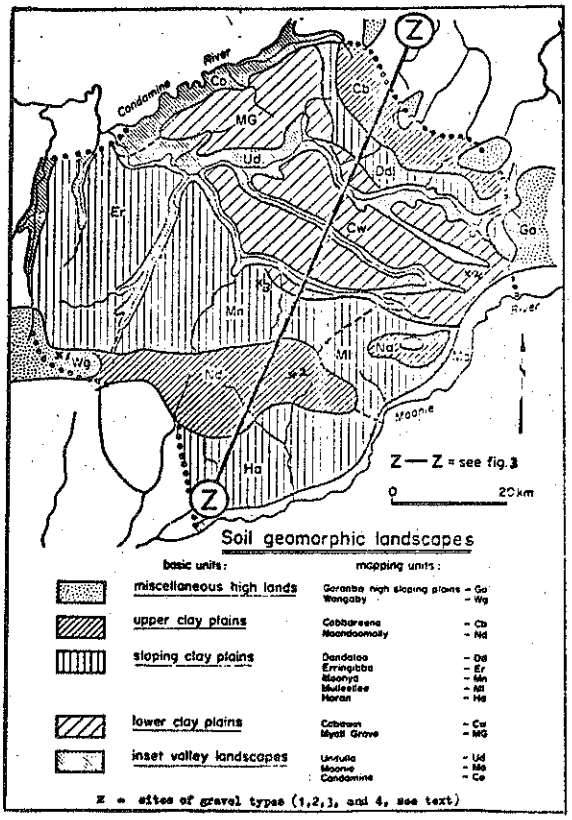

Fig. 2 LANDSCAPE SUBDIVISIONS OF THE TARA AREA

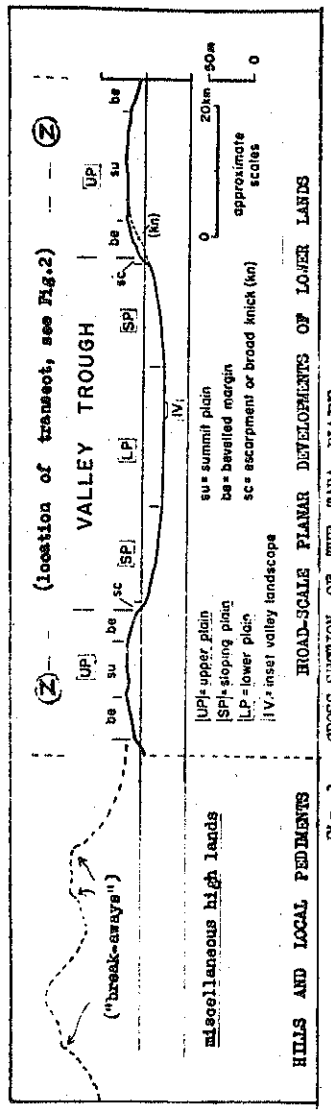

Fig. 3 CROSS-SECTION OF THE TARA PLAINS

THE PRODUCTION OF STORMFLOW RUNOFF FROM A BURNT, FORESTED HILLSLOPE NEAR WARBURTON, VICTORIA.

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ABSTRACT

This paper discusses hillslope hydrologic processes influencing the production of stormflow runoff from a burnt hillslope near Warburton, Victoria. Rainfall, throughfall, water-table behaviour and overland flow were measured automatically and continually over a 14-week period during Winter, 1983. It was found that despite reductions in canopy cover, elimination of the litter and ground vegetation layers, and the concomitant increases in ground receipts of rainfall, overland flow did not occur. Two explanations are advanced to account for this result. First, rainfall intensities in the area are far below the infiltration capacities of the hillslope soils. Second, hillslopes of the area are remarkably well drained, so the ability of the water-table to surface on the hillslope is limited both in space and time. As a variable source area schema of stormflow runoff production cannot apply to the study hillslope, a subsurface flow model is proposed to account for local stormflow generation. This finding supports results derived from a similar study situated in undisturbed forest from the nearby Cropper Creek catchment (Bren and Turner, 1979).

INTRODUCTION

Research into the hydrologic effects of land modification has been dominated by assessments of the impact of silvicultural and pastoral management practices. While significant efforts have been made to evaluate the ecological effects of forest fire (Ahlgren and Ahlgren, 1960; Kozlowski and Ahlgren, 1974), comparatively scant attention has been given to the impact of fire on hillslope hydrology. Australian studies by Langford (1976), O'Loughlin, Cheney and Burns (1982) and MacKay and Cornish (1982) have focused on the impact of fire on basin streamflow values and have tended to neglect hillslope processes relating to stormflow runoff production.

Highly combustible Eucalypt forest catchments provide 80 per cent of Melbourne's water supply (Langford and O'Shaughnessy, 1977a:1). Moreover, fire is frequently employed as a forest management tool in catchment areas to reduce combustible forest floor litter accumulations (Hodgson, 1970). Given these facts, it is surprising to note the lack of attention that has been given to the study of hillslope hydrologic processes on burnt hillslopes.

REGIONAL

The site chosen for study was on a burnt south-east facing hillslope adjacent to Smythes Creek in the Big Pats Creek drainage basin. The hillslope selected was of moderate gradient, having a slope of approximately 13°. The Big Pats Creek basin is located 6 kilometres south-east of Warburton township inn east-central Victoria. Large areas of the catchment were burnt during the "Ash Wednesday" bushfires of March, 1983.

As Warburton is situated at the foothills of the first major mountain range encountering the prevailing westerly and south-westerly maritime wind systms, a strong relationship results between rainfall and elevation in the area (Langford and O'Shaughnessy, 1977a:18). The long-term average rainfall for Warburton is around 1350 mm per year, though marginally higher totals could be expected in the Big Pats Creek catchment owing to its greater elevation. Rainfall intensities are generally low to moderate. when high-intensity rainfalls do occur, they are usually the result of summer-time convective cloud-bursts.

The unburnt vegetation of the area may be best classed as mixed species eucalypt open forest, canopy height being 25-30 metres and protective foliage cover about 70 per cent. the canopy layer is composed principally of E. radiata (narrow-leaved peppermint) and E. obliqua (stringybark) and is believed to be 1939 regrowth. the midstorey is rather dense and varied. Taller components include Blackwood and Silver Wattle rising above the smaller midstorey trees, Musk and Hazel. Dense assemblages of Cyathea australia (rough tree fern) and Dicksonia antarctica (soft tree fern) occur in the wetter gully areas. The groundlayer composition is clearly divided between wet and dry areas. In the drier upslope areas, a type of climbing wire-grass provides a platform for a thick layer of leaf and bark fall. In the moist downslope areas, the fern species Blechnum nudum (fishbone water fern) and Blechnum procerum (hard water fern) dominate the groundlayer.

Gully areas were lightly scorched while upslope areas were moderately to intensely burnt. Gully areas were less affected as they were moister, had less litter accumulation to fuel the fire and were often skipped by leaping flames. Fire damage increased upslope as the advancing fire was continually supplied with fuel and was fanned by upslope winds. Damage to vegetation included destruction of the Eucalyptus crowns, though not the death of the trees. In the upslope area, the midstorey and groundlayer were eliminated. In the gully area, the midstorey crown was lightly scorched, but the trees remained intact. Some groundlayer and tree ferns also survived and are now actively regenerating. Projective foliage cover of the canopy layer was reduced to 10 per cent following the fire. Forest floor litter was entirely consumed, thus baring the soil surface.

The soils of the study site may be described as brown, friable porous earths. According to the Northcote et. al. (1975) classification, they are brown, rough-ped earths (soil type Gn 4.31). These soils are quite deep and display diffuse, yet noticeable horizon differentiation. The surface of the soil is highly permeable, a condition which has been enhanced by good aggregation provided by high litter supply and an abundance of biological macropores. the upper horizon is highly interlaced by tree roots and channels created by land yabbies.

METHOD

Rainfall, throughfall, water-table behaviour and overland flow were recorded automatically and continuously over the 14-week study period. Fiture 1 details the field instrument layout at the hillslope study site.

Rainfall amount and intensity were measured automatically using an 8" diameter RIMCO tipping-bucket rainguage. Throughfall rates were measured in the upslope and downslope reaches of the study hillslope. Throughfall was collected in troughs which drained into RIMCO guages, thus allowing

both volume and intensity values to be computed. Water-table height was observed at four sites, spaced evenly along the hillslope transect (see Figure 1). Four observation wells were augered out and lined with perforated 80mm PVC tubing, capped to prevent direct rainfall contributions. Wells A,B and C reached the fractured bedrock, 60-70cm below the surface, while well D was augered to a depth of 120cm. The most downslope well (well A) was monitored automatically using a water-height sensor, while the other wells were observed manually during periodic visits to the site. Overland flow was measured at an upslope and downslope site using a paired-plot system. The plots were constructed by erecting walls of galvanised iron sheeting around demarcated areas to contain given rainfall contributions. The walls were pressed lightly into the ground to a depth of about 15cm and stood approximately 30cm above the ground to prevent wash-in. Plot area averaged about $29m^2$ and water was again channelled into RIMCO tipping-bucket gauges.

These variables were recorded automatically and continuously on a multichannel, magnetic-tape data-logger (MEMODYNE 2821, low power, 16 channel). Input signals were digitised and written down to standard digital cassettes which were readable by TEXAS INSTRUMENTS SILENT 700 and equivalent terminals. In order to minimise the bulk of data input, the recording system was operated on a periodic scan bases. An electronic quartz clock was programmed to switch the system on every ten minutes to scan the channels linked to the field instruments. This arrangement also saved valuable power and tape space. Inter-scan signals from the RIMCO gauges were stored on small memory interfaces attached to each tipping-bucket unit. A full description of the operation of the recorder interface is given by Dunkerley (1984).

RESULTS

Twenty-five storm events were selected from the data for analysis. A storm event is here defined as simply any sequence of rainfall interrupted by no more than two hours of no rainfall.

While monthly rainfall for the area was above average during the study period, rainfall intensities were quite low. All storm events recorded has return periods of less than one year. Moreover, the most intense rainfalls were shortlived, with intensities exceeding 12 mm/hr rarely lasting more than 20 minutes. Results emerging from this study must therefore be viewed in the context of the low intensity rainfalls prevalent throughout the study period.

Upslope throughfall rates were calculated for all 25 storm events analysed, though equipment failure at the downslope site meant that only 9 storm events could be used to assess throughfall contributions to the gully area. Upslope throughfall varied between 73.1% and 97.5% of gross rainfall measured in the open. Storm events of less than 5 mm were associated with upslope throughfall rates averaging 81.5%. Events measuring between 5 and 10 mm and those exceeding 10 mm were associated with upslope throughfall rates of 87% and 90.5% respectively. The level of explanation of upslope throughfall by rainfall (via the regression line) was 99.4%. Lesser throughfall receipts were measured in the gully reach of the hillslope. Though the sample size was small from the downslope site, regressed values indicated an explanation level of 95.9%. Rainfalls of less than 5 mm produced an average downslope throughfall rate of 60% while the comparative figure for events exceeding 5 mm was 77%.

Results from the automatically monitored well in the gully area provided a clear picture of water-table behaviour both between and during storm events. Discussion of the results of the water-table observation study is confined to values recorded at the downslope well (well A). There was no indication of water-table entry into wells C and D and only on two occasions did the phreatic surface enter well B to a maximum height of 40 cm below the surface.

From these results, four main features of water-table behaviour at the study site may be noted. First, water-table height was remarkably constant during low rainfalls and extended dry periods of up to five days (see Figure 2). At well A, the phreatic surface had a base height of between 0.3 and 0.4 m below the ground surface. Second, water-table response to major rainfall was very sluggish during periods of dry antecedent soil moisture conditions (see Figure 3). third, water-table response during periods of wet antecedent soil moisture conditions was rapid. During such conditions, surfacing of the phreatic surface was observed on five separate occsions (see Figure 4). Delays between peak rainfall and peak water-table height during these conditions were usually of the order of only one to three hours. A final feature of water-table behaviour noted the study site was the rapidity of water-table recession after conclusion of rainfall. This is clearly illustrated in Figure 4, where it can be noted that saturation of the soil profile rarely persists for more than a few hours after major rainfalls.

Small volumes of water flow (0.5 1/10 mins) were frequently recorded at the plot outlets during storm events, though such small quantities may be attributed to rainsplash near the plot outlets rather than to overland flow generated in the proper sense. Larger volumes were produced on only one occasion, during storm event 23. The production of overland flow during this event was restricted to the downslope plot and was associated with wet antecedent soil moisture conditions, a surfacing water-table and the highest rainfall intensity recorded at the site during the study period (32.4 mm/hr). A rainfall event of 5.4 mm between 1800 and 1810 hrs on 3/9/83 produced 1.5 litres of overland flow fromt he downslope plot. This voluem is still extremely small, representing less than 0.1% of potential runoff from an impermeable surface of equivalent area. It is argued that this volume can also be attributed to rainsplash generated near the plot outlets. As such, it may be concluded that no significant overland flow was generated from the experimental plots during the study period.

DISCUSSION

The two outstanding results of the throughfall process study were the high rates of throughfall measured at the upslope site and the relatively low rates measured at the downslope site. Langford and O'Shaughnessy (1977b) proposed a mean throughfall rate of 75.5±1.4% for undisturbed forest of similar type. Mean upslope throughfall at the Smythes Creek study site (86.5%) is therefore 11% greater and mean downslope rate (67.7%) approximately 7% lower than the Langford and O'Shaughnessy general value. Upslope increases are obviously explained by reductions in canopy leaf area and the elimination of midstorey intercepting surfaces. It is surprising to note however, that despite the mild effects of fire in the gully area, downslope throughfall rate is in fact lower than the general value of 75.4%. This is probably best explained by the fact that the general value proposed by Langford and O'Shaughnessy (1977b) does not adequately take into account variations in forest structure.

These results present a number of important issues. In predicting the hydrologic impact of forest fire, it is fair to expect that forest structural damage is likely to be unevenly distributed, with effects being most pronounced in upslope areas and minimised in moister, sheltered gully areas. This occurrence has obvious implications for runoff generation in that sheltered gully areas are often potential source areas of runoff generation. It is important that when measuring throughfall rates in both undisturbed and burnt forest environments, careful distinction be drawn between upslope and gully areas.

Litter and ground vegetation re-establishment at the study site has been remarkably rapid. Leaf fall from the scorched canopy had produced a light blanket of ground litter within two months after the fire. As early as five months following the event, eucalypts had shedded large amounts of bark which formed thick localised mats over the ground. The rapid and widespread invasion of bracken (Pteridium) and grasses, and profuse eucalypt seedling emergence have served to quickly re-establish ground cover. In essence, soil exposure at such sites is limited to the first two to six months following forest fire.

Although the available evidence of water-table dynamism is restricted to the downslope well (well A), it is fair to conclude that the ability of the water-table to surface on the study hillslope is strictly limited both in time and space. Results indicate that such hillslopes are extremely well drained and that expansions of saturated wedges such as depicted in the variable source area model of Dunne (1978), are unlikely to occur. As such, it is argued that the subsurface model of stormflow runoff production is more applicable to this particular environment.

Results from the runoff process study are consistent with those which emerged from a study by Bren and Turner (1979), based in an undisturbed environment of similar type in the nearby Cropper Creek Hydrologic Project Area of the Forests Commission, Victoria. Measurements of overland flow in this study indicated that less than 5% of incident rainfall reached the stream as overland flow, and that on most occasions this value was below 1%. The authors attributed these quantities to raindrop splash generated near the plot outlets on the premise that infiltration capacities of the hillslope soils exceed most incident rainfall intensities in the area.

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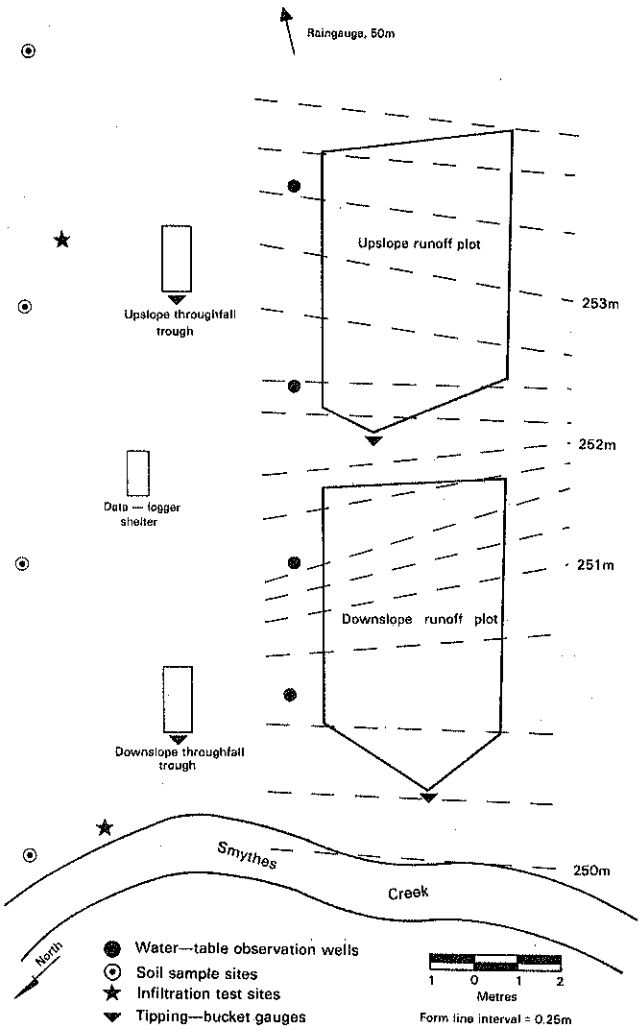
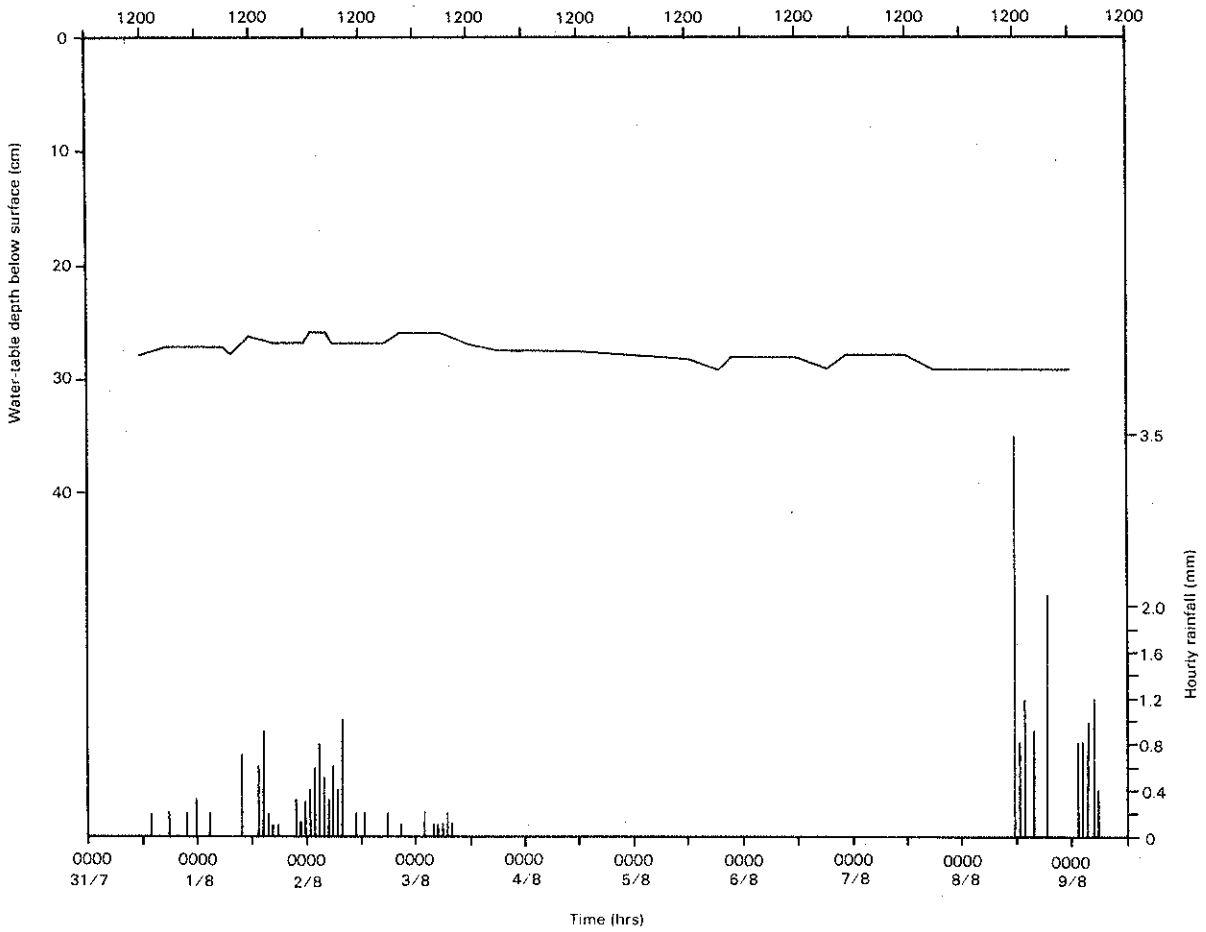


Fig. 1 Hillslope study site layout

Fig. 2 Water-table behaviour and rainfall 31/7-9/8



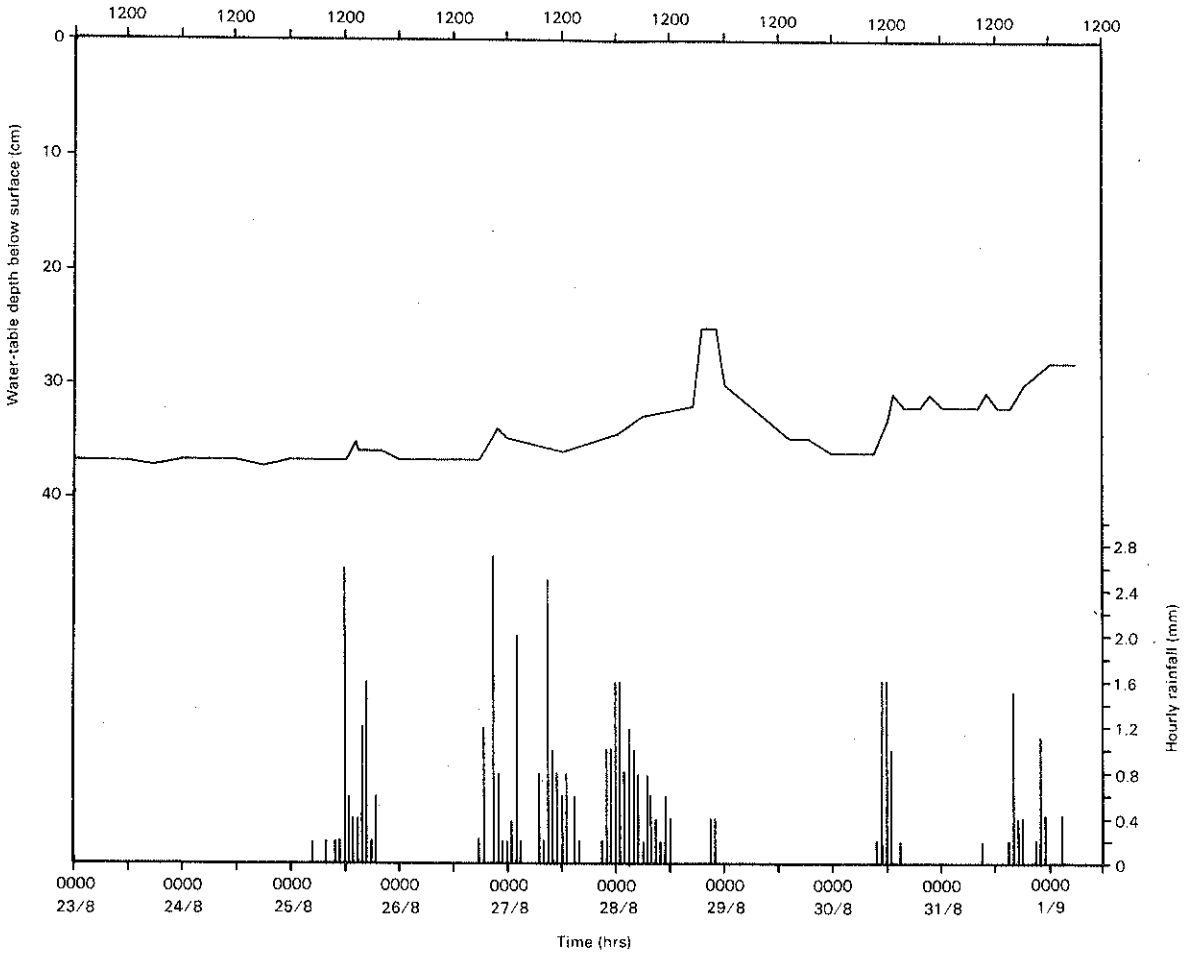


Fig. 3 Water-table behaviour and rainfall, 23/8-1/9

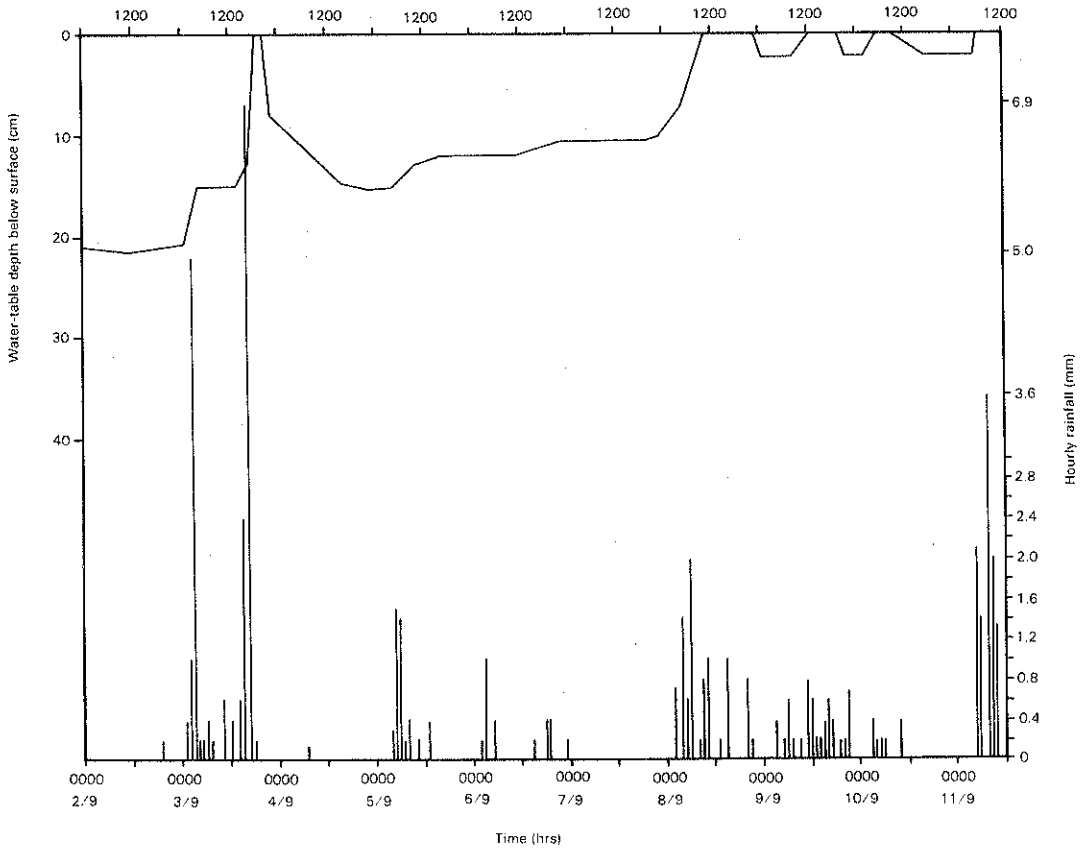


Fig. 4 Water-table behaviour and rainfall, 2/9-11/9

LANDSCAPE AND SOIL DEVELOPMENT IN THE COBAR-TILPA DISTRICT, N.S.W.

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ABSTRACT

Geomorphic-soil relationships along a transect through central western New South Wales near Cobar are described and hypotheses for landscape development are made. The study highlighted the complexity of these relationships and the very common recurrence of polygenetic soil profiles.

INTRODUCTION

The excavation of a two-metre deep trench in 1974-75 for the laying of the Moomba-Sydney Natural Gas Pipeline provided a unique opportunity to describe and study soils, geomorphology and vegetation and their inter-relationships along a transect across central New South Wales. Such a study was made by the Soil Conservation Service of N.S.W., the route being arbitrarily divided into four sections.

Basic soils and vegetation information for the western three sections, from the Queensland border to Burcher has been published (Lawrie and Stanley 1980, Walker 1980, Cunningham 1980).

The information collected has already been used for the formulation of land system maps of the region (Walker 1983, Johnston and Milthorpe in prep., Harland and Walker, in prep.). All soil samples collected have been retained and are stored at Condobolin.

The present paper summarises soil and geomorphic relationships within section 2 extending from the Darling River near Tilpa to near Mt. Hope (figure 1).

For the purpose of this paper the transect has been divided into seven broad soil-landform units (figure 1).

(1) Uplands on Igneous Parent Material

These form the oldest and highest elements of the landscape studied. Based on the Silurian Mt. Hope Volcanics (Wynwood and Penshurst land systems) and Thule Granite (Warrowie land system), they represent the more resistant rocks of an originally higher surface, with the highest peaks up to 40 metres above the surrounding lowlands.

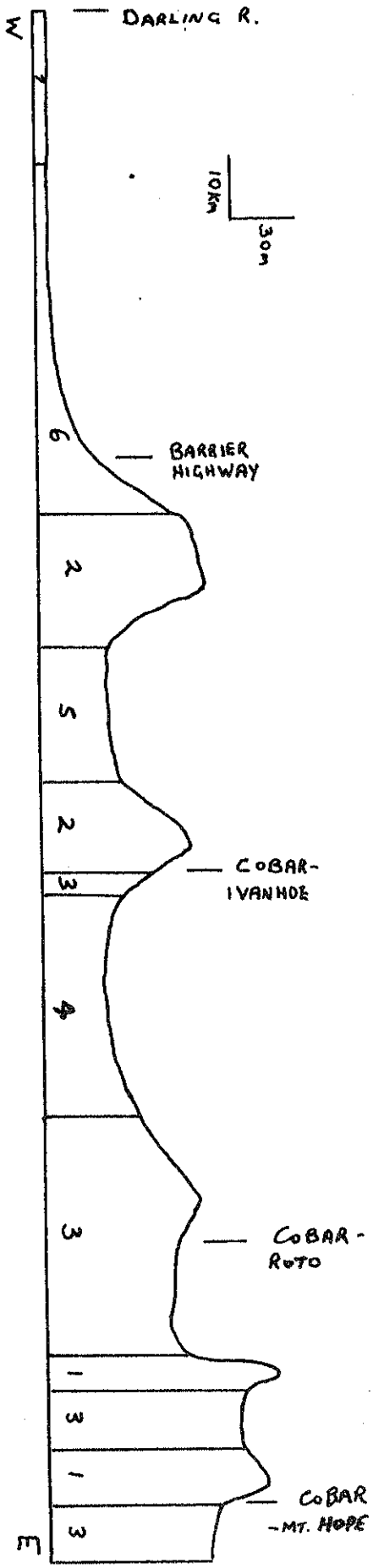

Figure 1. Landform units, relief and road and river crossings along study transect.

Erosion is a continuing process, particularly on steeper slopes, and rilling and gullying of lower slopes and minor drainage lines is minor to severe. Major drainage lines in Warrowie land system feature incised channels with heavy sediment loads of granitic sand and gravel which eventually form depositional fans on adjacent lowlands or lead to higher order ephemeral streams.

Soil textures reflect the coarse-textured parent rocks, but have moderate clay contents (sandy loam on ridge crests to clay loams and light clays on footslopes). Shallower upper slope soils have uniform texture profiles, but as depth increases downslope gradational red earths, massive red clays and various duplex soils (Gn 2.13, Uf 6.71, Dr 2, Northcote 1971) predominate. The upslope soils are considered to have formed in-situ, but considerable truncation and re-deposition have occurred on lower slopes and in drainage lines, as evidenced by gravel layers, earthy hardpans or abrupt changes in soil structure.

(2) Uplands on Sedimentary Parent Materials

These areas (Boulkra and Cottage land systems) are underlain by Devonian quartzites and sandstones, with crestal Tertiary outcrops in the vicinity of the Cobar-Ivanhoe road. Rock is exposed only on the higher ridges, which have relief to 100 metres, being overlain in lower positions by truncated lateritic soil profiles or colluvial deposits. These rocks also deeply underlie the adjacent plains. This landform is an erosion surface, with minor to moderate watersheeting and minor rilling the main present agents. Upper slope soils are

shallow and stony (Um 5.51). Ferruginous gravels cover the surface of most slopes, in part probably exposed and concentrated by erosion, and occur throughout upper layers or concentrated in bands with larger stones. The soils are gradational and relatively shallow (Gn 2.13) and overlie a matrix of calcrete, rock fragments and soil material, from a few centimetres to a metre in depth, in turn overlying rock.

Many profiles on lower slopes and in drainage lines show evidence of truncation and subsequent deposition of new soil parent material, in the form of sharp textural and/or pH changes and presence of well defined but narrow bands of stones or gravel.

Some pedological development, such as an increase in clay content and pH with depth, has occurred in these most recent soils.

On a ridge crest of Cottage land system an unexpectedly deep (1.7m) soil containing abundant carbonate and gypsum was encountered. This site is surmised to be a relict aeolian accumulation, since underlying rocks are carbonate-free, as deducted by Wasson (1982) from a study site to the south.

(3) Lowlands

These slightly undulating areas (Yackerboon, Taringa, Romani and Wilsons Tank land systems) have moderately deep soils, with rock rarely exposed and often more than two metres below the surface. However gravel and stone layers often occur. These soils have been formed partly in situ, but mostly inalluvial or colluvial deposits originating from adjacent and more distant uplands. Again there is ample evidence of truncation and subsequent redesposition. The soils have developed gradational texture profiles (Gn 2.13), with loamy topsoils and nodular carbonate usually present in the deep subsoil. This carbonate increases westward and forms prominent layers in the western unit. Many of these earths are abruptly underlain by structured blocky clays. "Duplex" profiles also occur, associated with more sloping areas and in some mallee sand plains. These appear to be polygenetic rather than formed by clay illuviation. Shallow loamy soils (Um 5.51) occur on higher ridges.

There is evidence in this unit of channel incision into the carbonate layer, with subsequent infilling with a carbonate-free soil material much more sandy than adjacent topsoils. This indicates occurrence of significant erosion events within the Quaternary period.

(4) Dunefield

A dunefield of parallel linear east-west trending dunes supporting mallee (*Eucalyptus* spp.) vegetation (Bindi land system) occurs in a basin between the two major upland areas. This dunefield abuts, along most of its eastern boundary, a north-south trending range of Devonian sandstones, quartzites and conglomerates. The soils of the dunefields show morphological similarity to those on and adjacent to these ranges and associated valleys and, like similar areas to the north (Walker 1982), are postulated to have

developed from alluvial material originating from these areas, with subsequent reworking by wind and possible aeolian additions.

Dune soils are uniform in texture in the upper layers and throughout higher dune profiles (Uc 5.11), but lower dunes and rises show an abrupt change at depth to a sandy clay loam containing some carbonate.

The dune-building mechanism is not known.

Swale soils vary according to depth of surface sand and may be duplex (polygenetic), gradational or uniform, and are usually calcareous at depth. The origin of this carbonate is not known, since the parent rock material is carbonate free (Wasson 1982). Wasson (1976, 1982) and Jessup (1961) have implicated aeolian accession in this process.

Along the western margin of the field, and associated with a former drainage system, soils are calcareous brown earths (Gc 1.22). Here dunes tend north-south, parallel to the direction of drainage, and are probably lunettes of small relict lakes or source-bordering dunes. This increased carbonate might be attributed to highly leached rocks in the upper catchment (unit 2), aeolian accession or a former marine intrusion.

(5) Creek Floodplains

These are level areas of alluvium apparently derived from south-western drainage from the Cobar Pediplain and to a lesser extent coarser sediments from Devonian sandstones to the north and from granite to the east, and border a few major incised channels (Mulchara and Kaleno land systems). The sediments are in general considerably finer than those of the dunefield, reflecting the finer textured rocks of the Pediplain. Deep loam to clay loam soils with gradational texture profiles (Gn 2.13, Gn 2.23) predominate, though duplex soils (Dr) and sandy rises (Uc 5.11, Gn 1.13) border channels. Dunes rarely occur, except as stream levees, again a reflection of the finer textured parent material.

(6) Plains

An extensive plain (Manara land system), with small dunefields (Bell Vale land system) and a transitional sandplain (Korreo land system) occurs between the Darling floodplain complex and the Barrier Highway.

The plain appears to comprise Quaternary alluvium older than that of the Darling (Bowler et al, 1978), of medium to fine texture and containing significant carbonate, which sometimes extends to the surface. The origin of this alluvium is not known but may be derived at least in part from the uplands to the east or may be the eastern-most extension of farmer marine invasion. The origin of the carbonate here also remains a mystery. Similar country, with belah (Casuarina cristata) - rosewood (Heterodendrum oleifolium) vegetation occupies a large area to the north west, south and west, with carbonate content increasing and depth decreasing westward.

Soils are calcareous red earths (Gn 2.13) in this vicinity and calcareous brown earths (Gc 1.1, 1.2, 2.22) to the west. A few low dunes occur, mainly to the east of small relict lakes, and small fields of parallel linear east-west dunes occur near the eastern margin in areas of coarser sediments probably derived from the adjacent eastern uplands.

It also seems possible that surface stripping of surface material from the plain by westerly winds occurred after deposition, removing fines, leaving calcareous sub-soils and supplying sediment for extensive dunefields (Tiltagoona land system) to the east (Walker 1982).

The generally easterly movement of coarser sediment is clearly evidenced by accumulation along the western flanks of the main Devonian ranges and uplands in this region. This may be in the form of sand sheets which decrease in thickness upslope (Korreo and Lachlan Downs land systems) or disorientated circular or irregularly shaped dunes adjacent to or astride ridge slopes. The trench exposure clearly demonstrated the former east of the Barrier Highway.

(7) Darling Floodplain

To the west of this unit lies a plain of mainly duplex soils (Dunoak land system) which is transitional to the Darling floodplain. Soil sequences on the latter have been described by Bowler et al. (1978).

General

The survey revealed the complexity of the topographic-geologic-geomorphic-soil relationships of this semi-arid region. Morphological examination permitted the development of the hypotheses for landscape formation briefly outlined above but further study using micro-morphological and carbon dating techniques is required to validate them.

The continuous soil exposure along the trench highlighted the considerable variation that occurs over very short (one metre or so) distances in soil depth or depth to a carbonate layer and the discontinuity of soil layers. This clearly indicates the need for adequate sampling to satisfactorily describe and classify soils. The large number of polygenetic profiles encountered often caused problems with established soil classification schemes.

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CHANNEL CHANGES: ADJUSTMENTS? TO WHAT?

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ABSTRACT

In this paper an attempt is made to link ideas on channel-forming discharges and channel changes, as defined by others and as observed in the Nepean. Here there appear to have been four major shifts in natural regime since European settlement, as well as modifications imposed by human activity. Stability is promoted by bank and bed cohesion and instability by its loss. Factors influencing both are considered, prior to a discussion of potentially high change areas using a new Pickup classification of channels. Eight ways of identifying changes are also given.

INTRODUCTION

Channel changes were once regarded to be slow and related primarily to the lateral migration of meanders. There have now been enough observations to reveal that channel type, size and shape can all change, sometimes faily rapidly and in response to more than meander migration. Accelerated changes have often been 'blamed' on human impacts in the channel and catchment.

If channel size and shape are related to some dominant discharge, and if it can be demonstrated that this can change significantly through time, and not just fluctuate about a mean condition, channel change may represent an adjustment to new dominant levels. In this brief paper the main aim is to examine channel changes, factors influencing changes and the identification of change, as well as what they might mean. In N.S.W. the assumed uniformity between channel morphology and dominant discharges is often illusive, while elements of instability are very common in those systems where channel change has been investigated.

Channel-Forming Discharges: The Assumptions which evolved

Consistency in the frequency of bankfull discharges in rivers with active floodplains was noted by many workers (Wolman and Leopold 1957, Dury 1959, Woodyer 1968). There was implied a causal relationship between channel dimensions and morphology with particular levels of discharge. The morphological significance of bankfull discharge in a channel-forming flow was recognised and exploited by Dury (1961). In reviewing this area, Pickup and Rieger (1979) make the distinction that "the acceptance of bankfull discharge as dominant discharge therefore implies that the river is 'in regime'". This means that channel morphologies are fluctuating about a mean condition, associated with a regime in equilibrium.

Where rivers are not "in regime" non-equilibrium or unstable morphologies display a distinct trend, which is manifested in progressive channel changes. Pickup and Rieger (1979) offered a channel-response model to accommodate the complications of such regime instability.

While such contributions reveal the complications and offer alternative strategies for investigation, the old notions die hard in some quarters and for a long time the idea that bankfull discharge was the channel-forming stage was something of a cornerstone in geomorphic thinking.

$$Q_{cf} = Q_{bf} = Q_{dom} = Q_{1.58/2.33} \quad (1)$$

where Q_{cf} is channel-forming discharge, Q_{bf} is bankfull discharge, Q_{dom} is dominant discharge and $Q_{1.58/2.33}$ are respectively most probable and mean annual floods (Dury, 1969).

Dominant discharges were found to be different by Benson and Thomas (1966) and Pickup and Warner (1976), along with others. The latter authors suggested that work on the Cumberland Plain revealed:

$$Q_{me} \approx Q_{dom} < Q_{1.58} << Q_{bf} \quad (2)$$

and that

$$Q_{bf} \approx Q_{6-20} \quad (3)$$

where Q_{me} is the most effective discharge for the movement of bedload and Q_{6-20} are discharges in the range of 1 in 6 to 1 in 20 year floods. These findings were for shale bedload streams west of Sydney with very cohesive banks.

Schumm (summarised in 1977) added perimeter conditions in the form of the weighted silt-clay percentage index to quantify bed and bank cohesion. This meant that shape, as well as size, could be related to sediment and water discharges. Width-depth ratios (F) were inversely related to silt-clay percentages (M) in the form:

$$F = 255\ M^{-1.08} \quad (4)$$

Again these results were obtained from stable channels "in regime". It is interesting to note that NSW coastal river cross sections, all plot below this line (Warner, Sinclair and Ewing, 1975), while inland rivers of the Namoi-Gwydir plot above the line (Riley 1975). It is probable that most coastal sections were unstable, but this might not have been the case inland.

For increases and decreases in water ($Q^{\pm}$) and sediment ($Q_s^{\pm}$) discharges, Schumm derived four equations where the direction of change for width (w), depth (d), meander wave length (λ), water surface slope (s), sinuosity (P) and width-depth ratio (F) might be predicted. Hickin (1983) in reviewing such changes concluded: "clearly, even for this qualitative scheme, many of the changes are indeterminate,.....because the magnitude of opposed responses is

unspecified". Actual and theoretical changes were compared for both sandstone and shale reaches of the Nepean, for both natural and human-induced changes (Warner 1983).

On the other side of this discussion, channel changes, which may or may not be adjustments to changes in channel-forming discharges, have received much attention in recent years (see reviews in Gregory 1977a, Rhodes and Williams 1979, Pickup 1981, Park 1981 and Hickin 1983). These and factors influencing change can now be considered.

Channel Changes

Any alteration whether very temporary or longer lasting, which can be observed in the cross section below banktop in a channel reach, can be regarded as a channel change. It involves either a loss of material such as bank erosion or bed scour, or a gain, such as deposition on the bed or banks. It is best demonstrated by changes in successive surveys, especially before and after a specific event. In practice, however, it usually represents net change over a longer period of time, eg. 1900 to 1982/3 on the Nepean (Warner 1984). If a system were stable or "in regime" such changes would not be great, representing "excursions from equilibrium conditions" (Hickin 1983). Within a shorter period of time this has been demonstrated for part of the Nepean and seemed to be related to the incidence and frequency of flooding (Warner 1983).

In the past, many changes have been seen as reestablishing equilibrium. For instance, a flood might cause severe erosion while later events might promote some recovery. Now it is evident that some changes are fairly rapid and indicate a trend, rather than fluctuations about equilibrium. This can be attributable to major shifts in regime and to changes imposed by human action in the catchment and channel. However, it is possible to regard such changes as part of longer, more violent departures from equilibrium (Pickup and Rieger 1979, Abrahams and Cull 1979). "Extreme events...may cause significant departures from equilibrium channel morphology that may persist for long periods of time" (Hickin 1983). He cites the well-known example from Cimarron Creek (Schumm and Lichty 1963). In such cases there is a "flood-dominated channel morphology" involving non-linearities in adjustment and what have been called complex channel responses (Hickin 1983).

This kind of situation seems to have prevailed in many NSW coastal streams with very erratic and changing regimes. In these rivers, channel capacity at any one time seems to be very much a function of the most recent batch of floods. In the 1950s particularly channels were severely modified with the passage of many catastrophic floods (Henry 1977, Abrahams and Cull 1979, Pickup 1976, Erskine and Melville 1983 and Warner 1972, 1983 and 1984). This period up to the late 1970s is now recognised as a secular change in climate (Cornish 1977, Pickup 1976, Riley 1981). The period 1900-1945 might have been an intervening quieter stage, where competence levels were lower and where channel capacities decreased.

The history which is being worked out for the Nepean seems to indicate these long period changes, although data for the first century of occupance are scarce. In 1900 there was a detailed survey above Penrith involving 46 cross sections. This was taken just

after the "devastating" 1900 flood and was at the end of a period of big floods which were reasonably frequent. This began in 1867 with what is still the largest flood on record (1900 is ranked 2). After this period of flood domination, the channel was very wide and shallow, with a slug of sand in the throat of the Nepean gorge (Warner 1984). During the succeeding period 1900-1949, there were few floods and, according to air photographs taken in 1949, widths were greatly reduced. It is assumed that depths increased at this time. From 1949 to 1970, in the early part of the secular change, widths increased again over most of the 5 km reach. Also in parts, this continued until 1982/3, the latest detailed surveys. In this period, depths increased for two reasons: sediment starvation with the closure of the Warragamba Dam (1960), together with high over-dam flows, and the creation of another barrier near the end of the gorge. The Glenbrook delta formed in the 1943 flood, nearly filled the gorge floor and has remained there ever since (Warner 1983 and 1984).

In this example there appear to be four periods: (a) pre 1867: relatively quiet in terms of flood dominance; (b) 1867-1900: big and frequent floods; (c) 1900-1949: a second quiet phase; and (d) 1949-1978 (at least): another period of big and frequent floods. While these have probably given periods of flood-dominated morphologies and intervening 'healing' periods, relaxation times have involved several decades. Also recovery of equilibrium in the long term is somewhat doubtful because of "lasting changes" imposed by human actions, particularly since about 1900. The channel has been regulated with 10 or more riparian weirs, four dams on the Upper Nepean and the Warragamba Dam affecting 80% of the catchment above Penrith. Some 40 mill m^3 of sand and gravel have been removed from the bed and flood plain above Richmond. Additionally, land use changes, conservation of farm water and land, as well as urbanization, have had profound effects on water and sediment discharges. The main impacts of all of these activities are in shutting off sediment supplies by the dams. These, plus the high runoff, have caused bed degradation in the gorges and bank and bed erosion in the shale-alluvium reaches of the river (Warner 1983 and 1984).

The same kind of problem has been addressed in theoretical terms by Pickup and Rieger (1979) and for an arid river by Graf (1983). The latter was able to use 112 years of data to investigate flood-related changes in the Salt River, Arizona.

Factors Influencing Change

Channel changes occur following loss of perimeter cohesion and erosion with the passage of a flow event or with the deposition of part of the load. In practice, this is scour and fill with the passage of a hydrograph. Where there is no progressive gain or loss over a period of time, there may indeed be equilibrium or "in regime" conditions. In the following section, factors promoting and reducing cohesion are considered, prior to a discussion of susceptible channel types. Factors promoting perimeter cohesion:

1. perennial flow with no great extremes or regular regime (virtually unknown).

2. low ratios of flood/mean flows and high flood/mean flood (again virtually unknown (Pickup and Rieger 1979).

3. high bank coherence (not uncommon outside sandstone and granite areas: Pickup and Warner 1976).

4. high bank coherence promoted by vegetation and root mats (riparian vegetation was dense, but thresholds often breached by clearance and flood erosion).

5. bed-armouring gravels (not common in sandstone and granite areas or in mobile zones (Pickup 1984)).

6. fine bed materials (more prevalent in western rivers).

7. steady through puts of water and sediment discharges (not common in flood-dominated regimes modified by regulation works).

8. bedrock perimeters (in gorges and common in entrenched valleys where they do constrain changes).

Factors which reduce cohesion are found to be more common:

1. intermittent and irregular flow regimes (big floods common in wet periods).

2. secular increases in precipitation increase flood magnitudes and frequency (Pickup 1976, Erskine and Bell 1982).

3. droughts are not uncommon (these lower water tables; cause cease to flow conditions; dry out banks and weaken cohesion; cause bank-protecting vegetation to die; promote bed colonisation and seedlings can become well established in long droughts, thereby increasing roughness, promoting sedimentation. All these lower bank resistance to the next flood).

4. high ratios for flood/mean flow and high flood/mean flood; high competence flows for scour and erosion. (Pickup and Rieger 1979).

5. low bank coherence in sand rivers or banks with interbedded sands (subject to rapid erosion, especially with loss of vegetation).

6. low bank coherence where vegetation in sparse or removed (there are records of clearance of vegetation both by floods and humans: thereafter bank erosion is pronounced).

7. sand-bed channels (wide and unstable: mobile zone (Pickup 1984, Erskine and Melville 1983)).

8. fine gravel channels (flow competence in Wiamamatta Shales was sufficient to give most effective discharges a frequency of 4 to 6 times a year (Pickup and Warner 1976)).

9. irregular supplies of sediments, either from the catchment or channel (high sediment delivery ratios can promote aggradation; low ratios can cause erosion, particularly following urbanization and conservation (Wolman 1967)).

Many coastal rivers would have reaches in this latter group and are potentially unstable. The Pickup (1984) classification of channel reaches in terms of mobility offers some means of distinguishing zones of potential vulnerability.

Sources zones are in bedrock areas where coarse or fine material is added directly to the channel, either regularly or irregularly. These loads are then transmitted by high competence flows.

The armoured zone is where gravels extend from the source areas into regions of softer rocks or pedimonts. While banks may be subject to erosion, beds are armoured against all but the highest flows. Equations developed by Graf (1983) indicate the frequency of such movements. In many coastal valleys source and armoured zones alternate until the plainlands are reached, where sediment delivery ratios are much lower. Armoured zones may be depleted below dams and where gravels are dredged for commercial uses as in the Nepean (Warner 1983 and 1984).

Armoured zones end abruptly (Pickup 1984). Below them are mobile zones where the bedload is finer and subject to movement by most flows. In rivers with regular regimes seasonal scour and fill will be common as in the Colorado (Leopold 1962), while in more erratic regimes there may be longer lasting changes. These have been investigated in the Macdonald by (Henry 1977, Erskine and Melville 1983) and in the Georges River by Warner and Pickup (1978).

Backwater zones exist where larger rivers tend to dominate smaller streams because, in the former, flood-plain levees accumulate faster thereby delaying tributary entrances and causing levee-dam swamps (Riley, Warner and Erskine 1984). These are areas of finer progressive sedimentation, eg Magela Creek to the East Alligator (Pickup, et al., 1983) or the Fly above the Strickland Junction (Pickup, et al., 1979).

Although sand is most readily moved, channel changes of all types are event based, relating to the passage of a hydrograph. In that event, water volumes are such that depths, gradients and velocities are reached where shear stresses are sufficient to overcome resistances in parts of the perimeter. Where more is removed than added, erosion results and,where more is added, there is deposition. The final part of this paper addresses problems of identification.

Identification of Channel Changes

Channel changes can be identified in many ways (Petts 1980), from casual observations over a long time to detailed surveys. Some have been 'discovered' accidentially, others by careful measurement, and others by looking for the consequences of changes elsewhere in the channel or catchment. Some means are briefly considered below:

1. gauge-site instability. Control shifts may fluctuate or trend in one direction, indicating progressive scour or accretion. This evidence has not be exploited to date.

2. Surveyed dimensions and position changes. Parts of channels have been surveyed for various reasons in the past. These can be resurveyed for comparisons to be made (Warner and Pickup 1978; Warner 1983 and 1984).

3. width changes from air photographs. Where stages are the same and banktops can be seen, channel width changes can be established. The weirs on the Nepean do give constant stages; elsewhere there can be problems with stage or state of tide or with riparian vegetation.

4. monitoring channel changes following catchment changes. Detailed surveys, the use of bank pins and other systems can be set up to establish controls before, during and after activities like deforestation (Rieger, Olive and Burgess 1979), urbanization (Leopold 1973), mining (Duggan 1981) and so on.

5. monitoring changes after channel modifications. Detailed surveys after tidal sand dredging (Warner and Pickup 1978), river sand dredging (Scholar 1974, Warner 1983), channel improvements (Bird 1980), the impacts of dams (Gregory, 1977b, Petts 1979, Warner 1983) may reveal impacts and recovery rates.

6. investigation of riparian vegetation. Bank tops are difficult to see through trees, but changes in vegetation can reveal channel changes in some cases.

7. ground checking using maps and air photographs. Catastrophic floods have been experienced in coastal valleys and impacts can be checked in the field (Henry 1977; Erskine and Melville 1983). Evidence of these events have survived 30-40 years in the Bellingen, Macleay and Moruya valleys.

8. defining thresholds for massive change. Graf (1983) was able to show that flows of specific magnitudes were responsible for flood-dominated change in an arid river. In this case knowledge of median-material size, slope, roughness, channel width and discharges enabled Graf (1983) to suggest that rupture of bed armouring would occur at 545 m^3/sec in the Salt River or at the 1 in 20 years flood magnitude.

Knowledge of this kind will enable the production of erosion-hazard maps or potential channel-instability maps which will aid catchment planners and managers.

CONCLUSIONS

Channel changes are of two kinds. They can be directly imposed by humans or they can occur following the passage of discharge events. The former may promote adjustments, such as those resulting from sediment starvation associated with a dam, or those associated with gradient steepening with channelization. Where rivers are "in regime", changes are such that they fluctuate about an equilibrium form and these represent adjustments or responses. Where systems are not "in regime", longer, more massive and progressive changes may occur. They are "complex responses" to major changes in water and sediment discharges. Rates of change, other modifications imposed by humans, and duration of trends which contain their own fluctuations are such that equilibrium conditions are

seldom achieved. Such conditions caused Hickin (1983) to state: "Perhaps most rivers are dominated by transient behaviour never fully adjusting to such events as major floods, climatic shifts, and those step-function effects known as river metamorphosis".

Individual levels of response may eventually be determined by careful and long studies of case examples, but it is by no means certain that levels of generalisation will be high. At this stage, it can be seen that the morphology of erratic rivers is at least partially a reflection of some inherited form and partially due to the passage of the last few major floods. These recently-acquired characteristics loom large in such cases and to suggest that size and shape are functions of some mystical, ever-changing, most probable or mean annual flood seems to be begging the question in this case.

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PATTERN AND PROCESS IN THE CONTINENTAL DUNEFIELDS OF AUSTRALIA

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Maps have been prepared of continental dune orientation, dune type and spacing using aerial photos, photo mosaics and planimetric maps. Each variable has been determined in cells of I5' x I5' on a Transverse Mercator Projection. Dune type has been assigned to one of six categories, while orientation has been determined from y-junctions in the case of linear dunes and overall shape in the case of parabolic dunes. Spacing has been measured, at right angles to dune orientation, along a I0km line centred in each cell and a mean ($\bar{s}$) calculated.

From these maps we see that the main dunefields (Great Sandy, Victoria, Mallee, Simpson/Strzelecki) are located in shallow basins. The well-known anticlockwise whorl of the dunefields is well displayed, and this most recent mapping has demonstrated that the whorl is closed in the west. Dunes previously have been recorded in the southern part of W.A. but their considerable extent is only now revealed. Dunes in western NSW have now been mapped over an area larger than hitherto recognized, and degraded dunes have been mapped as far north as I6^{o}S. It is now clear that dunes occur on about 39% of the surface area of the continent.

The dominant dune type is the linear narrow crested dune which occurs in all dunefields but dominates in the south and centre. Broad crested linears, often displaying multiple linear ridges on their crests, cover the next largest area but only dominate in the Great Sandy and Tanami/Sandover. Very short narrow crested linear dunes are almost confined to southern W.A. and western NSW, while parabolic and crescentic dunes are almost restricted to the Mallee. Network dunes are quantitatively unimportant and occur about the axis of the whorl. Confused dunes occur in all dunefields.

The map of mean spacing displays patterns that are of fundamental importance to our understanding of the way that dunefields evolve. The spacing between dunes varies within narrow limits about $\bar{s}$ over large areas and changes abruptly over distances of only a few kilometres (Wasson and Hyde, I983a). The map of $\bar{s}$ shows that the most closely spaced dunes occur in the interiors of the dunefields lying in basins, and the most widely spaced dunes occur on the margins. In other dunefields the pattern is not simple.

Wasson and Hyde (I983a) showed that there is a linear relationship between $\bar{s}$ and $\bar{h}$ (mean dune height) of linear dunes within statistically homogeneous blocks. The slopes of the least square lines vary between dunefields. Both $\bar{h}$ and $\bar{s}$ vary with equivalent (i.e. spread out) sand thickness (EST) according to the e^{x} exponential function (Wasson and Hyde, I983b). These morphometric relationships, along with the maintenance of s within strict limits over large areas, indicate that spacing is a fundamental variable of dunefield organization. An understanding of

the factors that control spacing is therefore likely to illuminate much of the geomorphic evolution of dunefields.

There are two likely groups of explanations of dune spacing and, therefore, size: I, control by atmospheric secondary flow (e.g. roller vortices); 2, control by sand transport processes.

No satisfactory model of the role of secondary flow has been formulated (see Hanna, I969, Leeder, I977). Roller vortices appear to lack the necessary coherence over large areas, apparently fail to generate adequate wind speeds, and have no identifiable role in dune initiation. Other secondary flows that could play a part are lee-waves and separation cells.

Once a dune grows by chance to a critical height it will induce lee-waves in its wake. If the lee-wave is strong enough it may be able to control the point beyond which another dune can survive. That is, the first descending wave has velocities higher than prevail at the crest of the upwind dune and so the swale is swept clean of sand as the wave trough shifts back and forth. Downwind of this 'cleaned' zone, a second dune can survive. This hypothesis does not account for dune initiation and also seems implausible because acceleration is insignificant in the wake of low hills (R. Scorer, pers. comm.). Moreover, lee-waves are unlikely behind hills less than c.40m high, and they occur under stably stratified atmospheric conditions when, in deserts, little sand is moved.

The hypothesis of spacing control by separation cells immediately in the lee of dunes (Twidale, I972), by the same process as that ascribed to lee-waves, is untenable because the cells extend downwind only about one dune height. Dune initiation is also not accounted for.

On current understanding, explanations of dune spacing that depend upon secondary flow are inadequate and so we turn to sand transport processes. Wasson and Hyde (I983b) showed that elemental dune types can be discriminated between by EST and a measure of the directional variability of sand moving winds. The second variable reflects the role of synoptic-scale circulation in dune formation but obviously winds acting at the meso and micro-scales are also important. A numerical simulation by Taylor and Gent (I974) of turbulent two-dimensional airflow over low hills has shown that the maximum surface pressure drop (Ph) at the crest of a hill is related to the geometry of the hill and upwind shear velocity (U_{*o}^2) in the following way:

$$\frac{Ph}{pU_{*o}^2} = \frac{h^{1.25}}{L^{0.25}Z_o}$$

where p is air density, h is hill height, L is downwind hill length and Z_o is upwind roughness. Shear and flow over a hill will be affected by Z_o so that dune h and spacing will increase with Z_o. This argument is set out in detail by Howard et alia (I977).

The map of $\bar{s}$ gives empirical support to this idealized relationship: $\bar{s}$ increases with the roughness of adjacent swales, so that closely spaced dunes occur in areas of fine-grained sandy substrates and widely spaced dunes lie on pebbly (gibber) surfaces. Mabbutt and Wooding (1983) have described the same relationship in the north western Simpson Desert where spacing changes by y-junctions in a downwind direction.

Further evidence of the importance of substrate type comes from measurements along seismic lines of dune h and s. In the Great Sandy, dunes are much lower and more closely spaced in topographic lows, such as relict drainage depressions, and more widely spaced on slightly higher ground. Aerial reconnaissance and examination of aerial photographs of this area has shown that bedrock and gravel often occur in swales where dunes are widely spaced, while swales between closely spaced dunes are certainly finer grained. These relationships also appear to apply in the Great Victoria Dunefield, judging from surveys along seismic lines, aerial photos and ground observations by G. Krieg (pers. comm.).

Transport rates of sand across swales and deposition rates on dunes vary with the texture and sorting of substrate sediments (Bagnold, 1941). Air passing over a sandy surface, at or above the threshold velocity for sand movement, erodes the sand at a rate governed by grain size, sorting, sand moisture content and surface roughness (e.g. vegetation cover). When sand is dry and free of vegetation, transport across well sorted fine sand is much slower than across a coarser and more poorly sorted sand. When sand-charged air passes onto a pebbly surface, the wind becomes undersatuarted with sand as the transport rate increases. Sand transport rate (Q) is given by the product of a coefficient C (dependent on, among other things, grain size and sorting) and $(V-V_t)^3$ where V is velocity at a particular time and V_t is the threshold velocity. The cubic power ensures that the effect of C on Q increases non-linearly so that Q increases 5 fold when a wind passes from a fine sandy surface to a pebbly surface.

A sand patch will grow into a dune if the rate of sandflow onto it is less than the rate of sand loss. On a pebbly surface there is very little <u>in situ</u> sand for dune construction, so most sand must come from upwind. It is plain that near the upwind boundary of a pebbly region the rapid increase of Q will lead to high EST but the frequency of undersaturated flows will probably result in widely spaced dunes. Further downwind $\bar{s}$ will increase as dunes terminate and so EST will decline as sand supply falls. The growth of large dunes on pebbly surfaces is aided by the behaviour of sand-charged air passing from a substrate onto a fine sandy surface (a sand patch or dune). Bagnold (1941) showed that the rate of deposition on the sandy surface increases as the textural difference between swale and dune surfaces increases. More sand will be deposited on dunes on a pebbly surface than on a sandy surface. The process of deposition is enhanced by greater turning of wind along sand/pebble boundaries compared with sand/sand boundaries.

Change of $\bar{s}$ often occurs abruptly, as noted earlier, and nowhere is it so marked than a few kilometres downwind of lakes and rivers. Large transverse dunes often lie along the edge of such sediment sources, with closely spaced linear dunes extending downwind from the crest of the transverse ridge. Within 10km the closely spaced dunes join in y-junctions and $\bar{s}$ increases. This change of $\bar{s}$ is often unaccompanied by a change of substrate texture, apparently in conflict with the argument developed above. This puzzle is not solved but it is tempting to think that the "form roughness" of the closely spaced dunes has effected a change in airflow which in turn has adjusted $\bar{h}$ and $\bar{s}$. The geometric relationship between Ph, Z_0, h and L, given earlier, suggests that such an adjustment is possible and illuminates a new direction for research into y-junctions (cf. Mabbutt and Wooding, 1983).

"Form roughness" created by dunes may be important in some areas but substrate type is a regionally more extensive control. The relationship between substrate type, Q and EST will vary between dunefields. The curves relating $\bar{h}$ to EST in Australia (Wasson and Hyde, 1983,b) are all of the e^x type (r^2 between 0.97 and 0.99) but the areas beneath the curves differ between dunefields; the greatest amount of sand being available in the Simpson/Strzelecki. Observations in this dunefield show that large stores of non-aeolian sand lie beneath the dunes so we conclude that sand supply is not equivalent to sand availability. Vegetation, sand moisture and grain size have all affected Q, reducing it most in the Great Sandy and Great Victoria dunefields. These two dunefields are presently less arid than the Simpson/Strezlecki and probably were so in past dune building episodes because of their greater proximity to the ocean. Thus, vegetation and sand moisture may have been more important in the Great Sandy and Great Victoria, so sand supply has always been low.

If transport rates have always been higher in presently hyper-arid dune-fields, by comparison with presently more humid dunefields, then it is possible that EST reflects, among other things, the presence or absence of long-distance sand transport by wind (Wasson and Hyde, in press). EST probably reflects the trap efficiency of dune types rather than being a determinant of dune type as implied by Wasson and Hyde (1983b). So star dunes trap almost all sand delivered to them, largely as a result of the converging wind regime in which these dunes form. Linear dunes trap sand more efficiently than transverse dunes which tend to rapidly pass sand downwind as they migrate, while barchans pass sand very quickly by streamer flow and by "turning over" (Lettau and Lettau, 1978).

Dune type is therefore seen as largely controlled by wind regime, and $\bar{s}$ and EST reflect sand supply and transmissivity of swale surfaces.

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WESTERN NEW SOUTH WALES ANALOGUE FOR PERMO-TRIASSIC ENVIRONMENTS OF BRITAIN

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The Permo-Triassic desert landscapes of north-west Europe have a modern analogue in the suite of late Quaternary fluviatile, eolian and lacustrine landforms that we have mapped between Ivanhoe and Cobar in semi-arid central-western New South Wales. We base this conclusion upon detailed geomorphic, stratigraphic and seismic studies of dunes, alluvial fans, ephemeral stream channels, and lake-lunette complexes in the Sandy Creek/Belarabon Ranges area, supplemented by a provisional radiocarbon chronology of major erosional and depositional events in the ~2 000 km^2 that we have surveyed so far.

SUBBASALTIC TOPOGRAPHY, BARRINGTON TOPS, N.S.W.

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Recently geomorphologists and geologists have shown renewed interest in the history of the eastern highlands. Contributors to the debate refer consistently to the work of Galloway (1967) in the Hunter Valley. Galloway had concluded that in the north of the region there was evidence for drainage reversal associated with a mid-Tertiary westward migration of the divide. Galloway also speculated about the degree of relief exhibited by the prebasaltic landsurface. He believed that the surface had a relief of no more than 400m.

The author is presently engaged in a detailed study of ancient drainage systems at Barrington Tops; located on the northeastern rim of the Hunter Valley. Basalts in the area have been dated to the Eocene (Wellman and McDougall, 1974). As a preliminary to palaeocurrent determinations on the subbasaltic alluvial sediments, the subbasaltic landsurface was contoured. The contouring was carried out by computer using the graphics package SURFACE II. Data on basal elevations of the basalt were drawn from air photos, field observations, Electricity Commission reports, and some restricted geological mapping carried out by two honours students from the School of Geology, UNSW. The resulting map, at a scale of 1:250,000, contour interval 50 metres, will be presented and discussed.

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ARID CLIMATE OR ROCK CONTROL: PROBLEMS IN THE INTERPRETATION OF SANDSTONE TERRAIN

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ABSTRACT

It has been repeatedly claimed that, even in a single rock type like sandstone, forms developed under arid climates can be readily distinguished from those developed under humid climates. The humid terrains used in these comparisons have, however, been subjected to climatic change. Observations in the southern Sydney Basin, which has suffered neither aridity nor widespread periglaciation, indicate that the supposedly diagnostic contrast in forms is largely due to variations in structure and lithology. Conclusions drawn for this region are extended to a critique of the methodology of much climatic geomorphology.

INTRODUCTION

The degree to which climate exerts control over the morphology of slopes cut in solid rock is still disputed. Budel (1980) assures us that "a natural system of relief formation has to be based exclusively on the exogenic processes which are closely governed by climate", and that other factors such as "petrovariance" are essentially "passive" (Budel, 1977, pp.13-16). Other climatic geomorphologists such as Louis (1979) or Tricart and Cailleux (1972) give much more emphasis to "azonal" influences, though Tricart and Cailleux do present an extremely detailed world-wide subdivision of morphogenetic zones. Yatsu (1966) warns us, however, that such classification is "merely a systematization of simple observations" which avoids the essential problems of landform development.

The general effect of arid climates on bedrock forms seems to have been the development of arrays of mainly steep and gentle slopes, with far fewer of moderate inclination (Mabbutt, 1977). Many authors have stretched this observation to the point where they claim that in a single rock type, say sandstone, dominantly angular slopes of arid regimes can be distinguished from the much more rounded and subdued forms of humid regimes. Indeed Tricart and Cailleux specifically chose sandstone terrains to argue the case for climatically controlled morphogenesis. In cases such as the "astonishing" wind-blasted terrain of Bembeche in northern Chad (Mainguet, 1972) the role of climate is beyond dispute. Its role in some other arid or semi-arid lands is hotly disputed; consider the argument as to whether slope forms on the Colorado Plateau have been controlled primarily by climate (Ahnert, 1960), structure (Bradley, 1963) or lithology (Oberlander, 1977).

Isolating the effects of climate has been made more difficult by the occurrence of supposedly humid forms in arid lands, and of arid forms in humid lands. This problem has faced us for 80 years or more, and is still unresolved. At the turn of the century Hettner (1903) pointed to the similarity between the cliff-lined valleys cut in the sandstones of Saxony, and the conyons and wadis of the Colorado and North Africa.

> A desert landscape, or should we say more accurately a landscape with desert forms lies in the middle of Germany! Have we therefore once had a desert climate which shaped these forms?

By answering in the affirmative Hettner prejudged the issue and effectively ruled out this region as a type example of humid terrain. It is the Vosges, where periglacial action is known to have subdued slopes, which has become the type example for comparison with arid lands (e.g. Tricart and Cailleux, 1972; Mainguet, 1972).

SYDNEY BASIN SANDSTONE TERRAIN

Clearly, what is needed for comparison with arid lands is a sandstone terrain which has been subjected neither to aridity nor to periglacial action. The southern part of the Sydney Basin is a case in point. Landforms here date back far into the Tertiary, but macro-fossil and palynological evidence shows that climates on the coastal margins of the uplands of southeastern Australia have been essentially humid since the Eocene (Martin, 1978; Kershaw and Sluiter, 1982). Dry phases did occur, especially during the Pleistocene (Singh, Opdyke and Bowler, 1981), but these were sub-humid rather arid, and postdated the forming of many of the rock slopes. Moreover, as the highest points in the southern part of the basin are under 1,000 m, Pleistocene periglacial influence was negligible.

One of the most striking features of this rugged landscape is the diversity of slopes forms which includes those supposedly diagnostic of arid conditions or those of humid conditions. Rounded slopes occur adjacent to angular ones; shallow upland valleys end abruptly at sheer rock faces; buttes rise steeply from undulating surfaces; tafoni (caverns) are as numerous as those reported from arid lands; and great cliff-lined escarpments, comparable to those which supposedly are limited to arid lands (Mainguet, 1972), abound.

Most of the processes which have shaped this terrain have already been described in the volume edited by Young and Nanson (1983) and can now be collated in a general model. Three points must be added to them. Firstly, convex forms are best developed where there are closely spaced bedding planes in the sandstone. Secondly, angular slopes seem to be the product of the mass failure of joint-bounded blocks. Thirdly, caverning is a major form of cliff disintegration. Structural and lithological variations have resulted in forms identical to those widely attributed to climate. Mainguet may therefore have been closer to the mark than she realised in saying that there is a unity of style in sandstone landscapes which repeats itself in all climatic zones. At least it seems that the only major difference between arid sandstone lands and this region is that here very broad pediments are rare, and the degree of dissection (and swamp development) on upland surfaces is greater.

METHODOLOGICAL PROBLEMS

Although this paper has dealt only with sandstone terrain, it points to more general problems in the methodology of climatic geomorphology, especially to the categorising of landforms according to climate.

1. The categories have been drawn up not in reference to any specifically defined problems, but to throw light on the general question of climatic influence though that question remains vague.

2. Categorising of this type may lead to Whitehead's fallacy of misplaced concreteness, in that we forget that the categories are just labels for a point of view.

3. The categorisation is based on the assumption that there is one control which, at least for a great number of landforms, overtops the rest. Yet all landforms can be looked at from a number of points of view and be categorised according to the viewpoint selected.

4. The classifications of climatic geomorphology have been derived from very biased sampling, especially because of the major climatic fluctuations in northern mid-latitudes. Troll's point that the Southern Hemisphere is not a mirror image of the Northern has been generally ignored.

5. The task is not to refine the categories (a la Tricart and Cailleux) but to formulate problems so that they are separately investigable.

The influence of climate cannot be denied, but, as Mabbutt (1977) emphasises, perhaps the best work in arid lands has combined the climatic with the morphostructural approach.

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THE AGE AND ORIGIN OF CHINESE TOWER KARST

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Evidence is presented to support the following points:

(a) Karst towers such as found in South China are time-transgressive landforms.

(b) Their initial form is inherited from a previous polygonal karst phase with conical residual hills.

(c) Their development is closely controlled by hydrological factors, footslope undercutting only occurring in areas liable to inundation.

(d) Significant uplift results in the reversion of tower karst to polygonal karst.

(e) Tower karst of lowland tropical to subtropical origin has been uplifted at least 2000m in parts of South China, and possibly to 4000-5000m in parts of Tibet.

Review of the international literature indicates that there are four main types of tower karst:

1. residual limestone hills protruding from a planed limestone surface veneered by alluvium;

2. residual hills emerging from limestone inliers in a planed surface cut across non-carbonate rock;

3. limestone hills protruding through an aggraded surface of clastic sediments that buries the underlying karst topography; and

4. isolated limestone towers rising from steeply sloping pedestal bases of various lithologies.

The first type is prominent in the areas investigated in China. The paper discusses its characteristics and provides evidence of the effect of minor base level changes. Paleomagnetic results from cave deposits in the towers near Guilin give no indication of reversals, indicating that sediments in the caves and constituting river terraces of similar altitude outside are less than 0.69 My old.

Uplifted karst of the first type was also examined on the Guizhou plateau, where veretbrate and plant remains suggest a moist and hot climate in the past. Pollen evidence from karst localities in Tibet point to more than 3000m of uplift, possibly since the Pliocene. Valley incision and knick-point retreat into uplifted tower karst is now leading to its rejuvenation.

TIDAL FLUX AND THE MOVEMENT OF PARTICULATE MATTER IN A MANGROVE BASIN, AUCKLAND, NEW ZEALAND

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ABSTRACT

Two methods which have been widely used in salt marsh systems to estimate material fluxes were applied to the study of tidal flux and the movement of particulate matter in Tuff Crater, an old explosion crater filled with mangroves, in the Waitemata Harbour, Auckland. The direct measurement method estimates discharge by determining mean velocity and cross-sectional area; the hypsometrically-based volumetric method is based on the relationship between volume of water stored in the basin and tidal stage, and approximates discharge from changes in stage. There are shown to be disparities between the two methods which effect not only the total estimate of suspended sediment movement, but more importantly the magnitude and direction of net material flux. The volumetric method appears inappropriate for determination of particulate matter budgets because of invalid underlying assumptions.

INTRODUCTION

Coastal wetlands, comprising both temperate salt marshes and tropical mangrove swamps, are generally areas of fine-grained sediment accumulation, considered to sustain a high rate of organic primary production which supports higher trophic levels in detrital-based marine food webs. Geological studies of these intertidal halophytic communities emphasise the import of sediments and the accretion of mud, while biological studies indicate a net export of organic production, of nutrients, particulate matter and floating detritus.

Salt marshes and mangrove swamps are flooded and drained by the tides via dendritic tidal creek systems. These creeks are characterised by bi-directional flows, but there are few data to indicate that they serve as avenues for net import of sediment and net export of organic matter. A number of studies have examined flow characteristics, and sources of error in the estimation of material fluxes through salt marsh creeks, especially along the eastern coast of the U.S. (Settlemyre and Gardner, 1977; Boon, 1978; Ward, 1981; Kjerfve *et al.*, 1981). However the role of tidal creeks in the movement of material in mangrove swamps has hardly received any attention.

This study examines the application of two techniques, which have been applied widely in attempts to estimate budgets of material in salt marshes, to a mangrove basin, Tuff Crater, in the Waitemata Harbour, Auckland, New Zealand. The work was undertaken as a part of a broader study of the productivity and function of a swamp of the mangrove, *Avicennia marina* var. *resinifera*, and the area chosen was in an old volcanic crater which facilitates the estimation of tidal flux.

AREA OF STUDY

Tuff Crater (36° 48'S, 174° 45'E) is an old explosion crater adjacent to Shoal Bay in the Waitemata Harbour. The crater walls have been breached and the basin has been inundated by the sea, infilled with marine sediments and the surface, of 21.6 ha, colonised by the mangrove Avicennia (Woodroffe, 1982). The mangrove-covered mudflat is inundated by the tide via a tidal creek which runs through the breach in the crater walls and flows beneath a culvert under the Northern Motorway (Figure 1). Tides are semidiurnal, with a tidal range in the Waitemata Harbour of 2.69 m at springs and 1.99 m at neaps. The culvert does not allow the full tidal volume to enter the basin at the highest tides and water levels at spring high tides are several centimetres lower in the basin than in Shoal Bay. This site is particularly suitable for a study of tidal flux because it is clearly defined by the crater walls, does not have a freshwater stream input, and has a narrow outlet, at the breach in the crater walls, through which all exchange takes place.

METHODS

Measurements of tidal flux were made from a shallow-draught dinghy moored in the centre of the tidal creek in the breach in the crater walls (Figure 1). An Ott current metre was used to gauge the velocity of flow at 0.2, 0.6 and 0.8 of water depth at 15 min. intervals. Stage was recorded on a stage board and cross-sectional area of the creek calculated accordingly. Discharge for 15 min. periods was estimated from mean velocity and cross-sectional area. It was assumed that on overbank tides exchange of water occurred in the creek and was negligible through the adjacent mangroves.

In addition the absolute elevations of more than 400 points on the mudflat surface and of 12 creek cross-sections were surveyed. These data were used to calculate the volume of water stored in the basin as tidal stage varied.

The concentration of suspended sediment was measured by filtering water samples on glass-fibre filter papers (Whatman GF/C). Water samples were taken at 15 min. intervals, integrated over the entire water depth, and filtered on a Buchner funnel. The dried filter papers were then ignited in a muffle furnace at 500°C in order to assess the organic fraction of the particulate matter. Total suspended sediment (T.S.S.), inorganic suspended sediment (I.S.S.) and organic suspended sediment (O.S.S.) were determined in this way.

RESULTS

Direct measurement of tidal flux by velocity-area methods

The tidal creek fills and the mudflat is inundated on the flood tide; at slack high water the velocity decreases and then the flow reverses and the basin empties on the ebb. At a stage equivalent to mean sea level (M.S.L.) there is water only in the deepest parts of the creek system and the entire basin dries at low tide except for a gradual persistent trickle from the bed of the creek.

The velocity of flow was found to be relatively constant with depth in the creek. Mean velocity was calculated from readings at the 3 depths and varies with the stage of the tide. On neap tides flows are below-bank and water does not inundate the swamp surface. Most tides result in over-bank flows, and velocity increases rapidly when the banks are overtopped on the flood limb. Similarly there is

a peak in velocity on the ebb limb as the water drains from the swamp surface. The peak velocity on the ebb consistently exceeded that on the flood, and occurred at a lower stage, on the 12 tidal cycles gauged. In addition peak velocity is greater the higher the maximum stage of the tide reached.

Total discharge over a tidal hemicycle was estimated by summation of values for 15 min. periods. An example of the discharge-time curve using this method is shown in Figure 2a for the tidal cycle on 8 November 1982. There are experimental errors inherent in measuring mean velocity, and on the 12 tidal cycles examined the disparity between the total discharge for the flood and for the ebb varied from <0.1% to 40%. On half the tides the error was less than 10%.

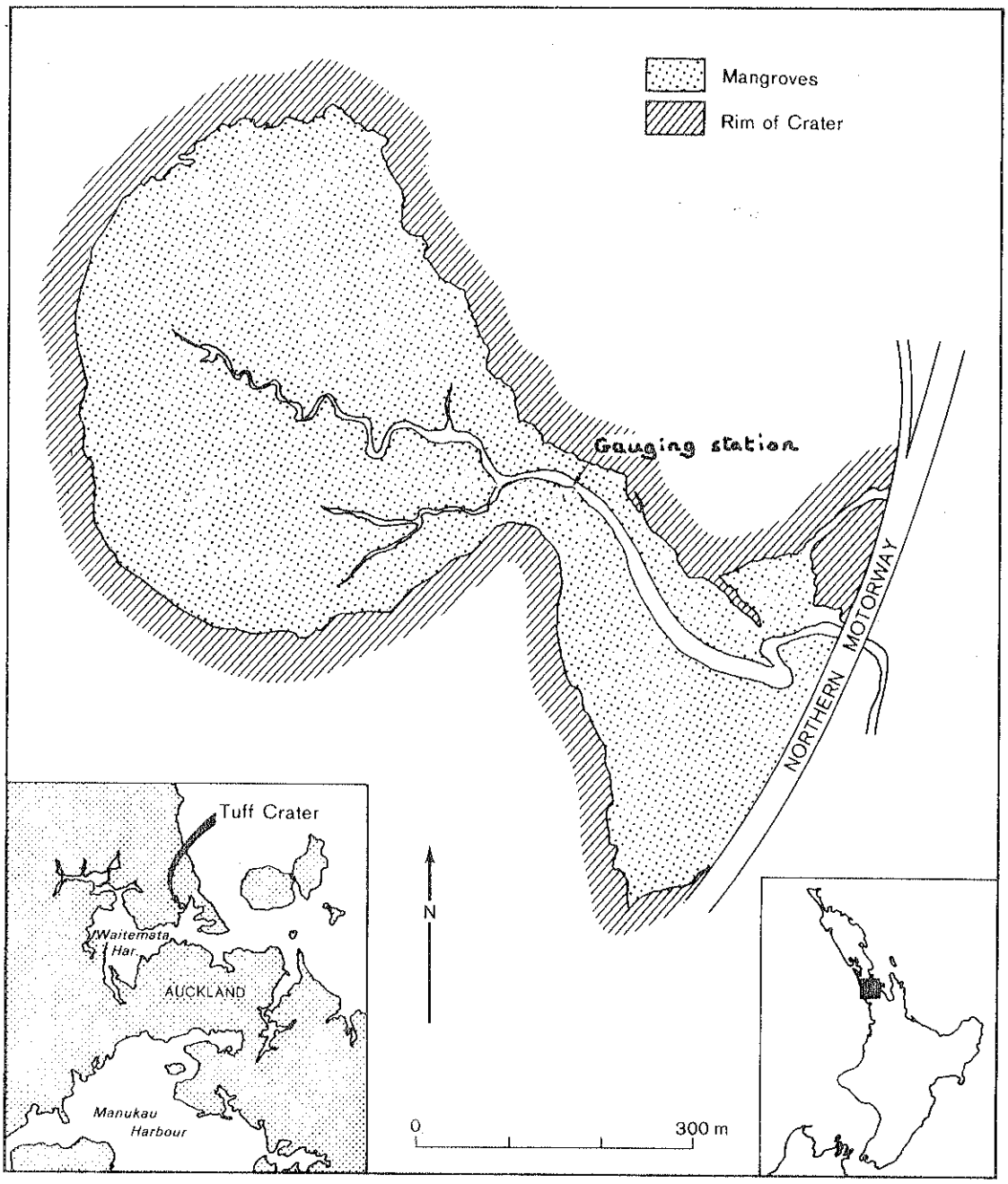

Figure 1. Tuff Crater, a mangrove-filled volcanic crater, its location, and the position of the gauging station.

Approximation of tidal flux by hypsometrically-based volumetric method

The detailed topographic control of Tuff Crater permits estimation of the volume of water stored in the basin at any tidal stage, assuming the water surface in the basin to be horizontal. At a stage of 1.00 m above M.S.L. water is contained almost entirely in the creeks (6285 m^3 in creeks and 2482 m^3 on the mudflat surface), whereas at 1.20 m above M.S.L., 35500 m^3 of water is stored, the majority over the mudflat surface.

Discharge for 15 min. periods was calculated for both tidal hemicycles using the volume-stage curve for the basin, and observations of stage taken at the gauging station at the beginning and end of the 15 min. period. This was done for the 12 tidal cycles for which direct measurements were made. The discharge-time curve using this hypsometrically-based volumetric method for the tidal cycle on 8 November 1982 is shown in Figure 2a.

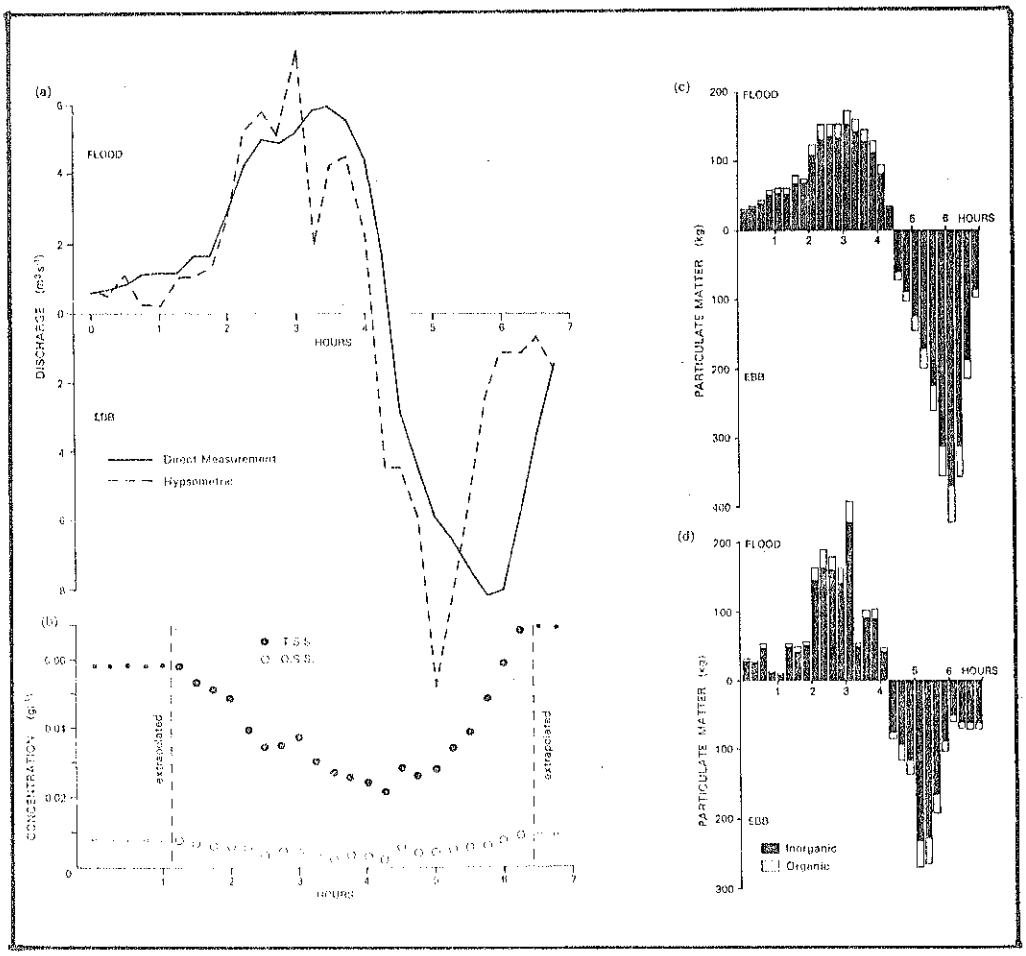

Figure 2. Tidal flux and particulate matter characteristics of a tidal cycle on 8 November 1982: a) Discharge-time curve for the tidal cycle measured by the direct measurement velocity-area method and the hypsometrically-based volumetric method. b) The concentration of total suspended sediment (T.S.S.) and organic suspended sediment (O.S.S.) over the tidal cycle. c) Particulate matter-time histogram using the direct measurement velocity-area method of estimating tidal flux. d) Particulate matter-time histogram using the hypsometrically-based volumetric method of estimating tidal flux.

Comparison of estimates of tidal flux using the two techniques

The discharge-time curves estimated using the two techniques show minor disparities but generally the shape and magnitude of the flood and ebb discharge are similar (Figure 2a). There are two differences observed on the 12 tidal cycles, which are of note: i) the volumetric approach predicts an earlier turn of the tide, and ii) the volumetric approach predicts an earlier peak discharge on the ebb tide.

Not only are these two differences observed on each tidal cycle but they were also shown to exceed measurement errors. The earlier turn of the tide predicted by the volumetric models occurs because the water movement is presumed to cease when the water level ceases to rise at high tide. In practice water continued to flow into the basin while the water level was static or even for the first centimetre of fall in stage.

The peak discharge observed on the ebb was as much as one hour later than that predicted by the volumetric model. This presumably results from inertial effects retaining water on the mudflats. The observed peak discharge occurs at a stage when water level is generally below the banks, and it appears that the water surface is no longer horizontal. Discrepancies between observed and modelled discharge have been observed on the ebb flows in salt marsh creeks (Boon, 1975; Healey et al., 1981), and it appears that the assumption underlying the volumetric technique that water surfaces are horizontal is unrealistic.

There are also disparities in the total discharges calculated by the two techniques. In all cases the direct measurement technique estimates exceed those of the volumetric (from <6% to >67%). It is probable that this is due in large measure to the fact that the direct measurement method does not make allowance for cross-sectional variation in velocity.

Particulate matter concentration and flux

The concentration of particulate matter varies over a tidal cycle with lowest concentrations observed at high slack water, and with concentrations rising rapidly on the ebb limb. Variation in T.S.S. and O.S.S. over one tidal cycle is shown in Figure 2b.

Total flux over a tidal hemicycle was estimated by summing estimates of particulate matter flux for 15 min. periods, the product of discharge and suspended sediment concentration. A particulate matter-time histogram was determined for both the direct measurement method (Figure 2c) and the hysometrically-based volumetric method (Figure 2d) of estimating tidal flux. These histograms indicate different rates of movement of suspended sediment. This difference is particularly marked on the ebb limb and occurs because, while the results of direct measurement indicate that peak discharge coincides with high suspended sediment concentrations, the earlier peak predicted by the volumetric method occurs when concentrations of T.S.S. and O.S.S. are much lower.

Particulate matter flux was calculated for 11 of the 12 tidal cycles for which flux was measured and results were similar to those in Figure 2. The estimate of volume of material moved was less using the volumetric method than the velocity-area method. This, in part, reflects the lower tidal flux estimates using that method. However while the volume of material averages 19% less on the flood using the volumetric estimate of tidal flux, that on the ebb averages

32% less. The covariance of suspended sediment concentration and discharge means that the displaced ebb peak discharge of the volumetric method coincides with lower concentrations than the actual peak and will thus underestimate the movement of material on the ebb. The estimates of net flux over 11 tidal cycles in Tuff Crater were consistently more positive (favouring import) using the volumetric method, than those by direct measurement.

CONCLUSION

Several studies of salt marsh creek systems have been undertaken to show the source of errors in the determination of material fluxes over tidal cycles, pointing out that where the net movement is determined as the difference between a large total flood and a large total ebb movement, the net value is generally of the same order of magnitude or smaller than the errors accumulated by measurement methods (Boon, 1978; Kjerfve *et al.*, 1981). In studies where net budgets, of sediment or organic carbon or nutrients, have been determined, tidal flux has been calculated either by direct measurement or by the volumetric method. This study of tidal flux in a mangrove basin, Tuff Crater, using both methods indicates not only that there are differences in the rates of movement calculated using the two methods, but, more importantly, that the magnitude and direction of the net flux may be dependent upon which method is used. This unique basin topography makes it particularly suitable for the study of flux and it is implied that the volumetric method is based on inappropriate assumptions and does not give a valid estimate of net material flux. Both methods have been used for the calculation of budgets in coastal wetland systems elsewhere, and the comparability of results must be questioned.

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