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GEOLOGICAL SURVEY BULLETIN 68

The Quaternary geology and glaciation of the King Valley

TASMANIA DEPARTMENT OF MINES

COVER PHOTOGRAPH

View looking southward down the King Valley from Marble Bluff. Dante Rivulet in foreground, King River with shingle banks in middle ground. Comstock Valley, with Dante outwash fan, in right foreground. Mt Owen, Thureau Hills, Mt Jukes and Mt Darwin in distance.

K. D. Corbett

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The Quaternary geology and glaciation of the King Valley

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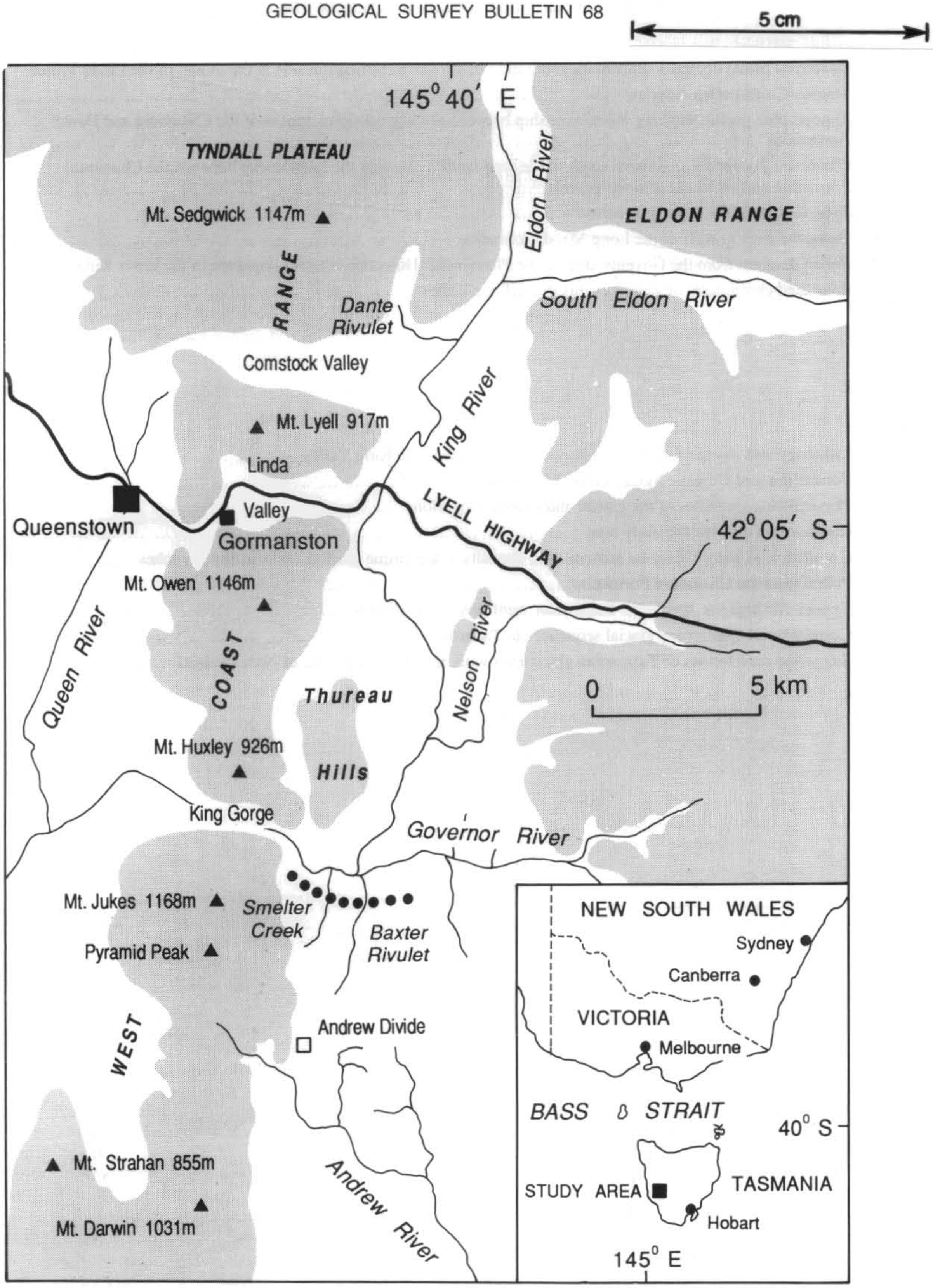


Figure 1. Location map showing the topography and drainage system of the King Valley. The shaded area is land over 400 m and the dotted line is the southern limit of Jurassic dolerite clasts in Quaternary sediments. The dolerite is derived from a sill that caps the Eldon Range and Mt Sedgwick.

ABSTRACT

Mapping, analysis and interpretation of Quaternary deposits in the King Valley, western Tasmania permits the identification of six climatic stages and has led to revision of the Quaternary stratigraphy of the West Coast Range. The oldest deposits in the valley predate glaciation, contain extinct pollen types and are probably of late Tertiary age. Overlying deposits of the Linda Glacial Stage are intensely chemically weathered and have a reversed detrital remanent magnetisation indicating deposition before 730 000 yr B.P. The highly weathered tills are overlain by organic deposits of the Regency Interglacial which show a transition from montane scrub rainforest to lowland temperate rainforest. Deposits formed during the later Moore Glacial Stage record advances of the King Valley Glacier and glaciers from the West Coast Range. A pollen-bearing fluvial deposit records an interstadial interval during the stage. On the basis of weathering rinds, amino acid dating and palaeomagnetism the deposits are estimated to have formed between 400 000 and 500 000 yr B.P. The Moore

Stage deposits are overlain by sediments formed during the Henty Glacial Stage. These deposits are believed to predate 130 000 yr B.P. They record multiple advances of the King Glacier and the development of a large interstadial lake. Deposits formed by ice advances during the last or Margaret Glacial Stage are restricted to a small area in the northern part of the valley. Although the most recent ice advance culminated about 19 000 yr B.P., evidence of older Margaret Glacial Stage deposits suggests that an early Last Glaciation ice advance may have occurred before 48 000 yr B.P. When combined with earlier studies, the recent work in the King Valley has provided one of the more complete records of Pleistocene glaciation in the Southern Hemisphere.

INTRODUCTION

This bulletin is concerned primarily with the Quaternary glacial deposits in the upper part of the King Valley from the confluence of the Eldon and South Eldon rivers to the entrance of the King River Gorge through the West Coast Range (fig. 1). During the Quaternary this area was a major centre of glaciation. A small ice cap formed on the Tyndall Plateau and flowed across the Lake Dora area and south along the King Valley as an outlet glacier. Numerous small cirque glaciers formed on the peaks of the West Coast Range and flowed east onto the floor of the valley.

For the most part, this bulletin is based on a Ph.D. study of the area recently completed by the senior author (Fitzsimons, 1988). Stratigraphic units mapped during the study have been incorporated into the 1:25 000 geological map of the Queenstown area published by the Department of Energy and Resources (Corbett *et al.*, 1989).

The King River hydro-electric scheme will inundate much of the King Valley in 1992, at which stage many of the deposits described herein will become inaccessible.

Physical setting

The King Valley lies in a belt of complexly faulted and folded Cambrian, Ordovician and Siluro-Devonian rocks (Corbett *et al.*, 1977). The eastern margin of the King Valley is bounded by faults that separate Siluro-Devonian sediments which crop out in the valley floor from Precambrian schists and quartzites. To the west, the valley is bounded by folded Cambrian volcanic rocks and by Ordovician conglomerates which form the West Coast Range. In the north the valley is bounded by the continuation of the West Coast Range and the Eldon Range which consists of Siluro-Devonian sediments, Permian sediments and Jurassic dolerite.

The broad elements of the King Valley landscape are related mainly to variations in lithology and to fault and fold structures. Most of the valleys are structurally controlled. The broad U-shaped, flat-bottomed King Valley is part of the King Synclinorium that is bounded to the west by the West Coast Range and to the east by a faulted contact with Precambrian rocks. The central part of the King Synclinorium is occupied by the less resistant Siluro-Devonian Eldon Group sediments and the Ordovician Gordon Limestone. The Linda and Comstock valleys are also controlled by faults. The courses of the Princess, Nelson and South Eldon rivers are controlled by local variations in lithology and the steep dip of the Siluro-Devonian rocks.

Three distinct lithological provenances can be recognised in the drifts of the King Valley. These are the Jurassic dolerites and Permian sediments of the Eldon Range, the Cambrian volcanic and Ordovician conglomerate rocks of the West Coast Range, and the Siluro-Devonian sediments derived from the valley floor (table 1). The limited outcrops of both Permian sediments and Jurassic dolerite in the source areas for the King Glacier make them useful indicators of the direction of movement and extent attained by southward flowing glaciers in the valley. Similarly, the absence of these rock types from sediments dominated by Owen Conglomerate and Cambrian Mt Read Volcanics can be used as evidence for derivation from the West Coast Range. The Eldon Group rocks derived from the valley floor constitute a significant component in almost all Quaternary sediments and therefore do not assist in defining the direction of ice movement or extent of glaciation.

Table 1. LITHOLOGY AND SOURCE OF ROCKS IN QUATERNARY SEDIMENTS IN THE KING VALLEY

Rock type	*Source*
Jurassic Dolerite	Eldon Range (erratic)
Permian sediments	Eldon Range (erratic)
Siluro-Devonian sediments	Valley floor
Ordovician conglomerate	West Coast Range
Cambrian volcanics	West Coast Range

A major feature of the drainage that does not appear to be geologically controlled is the King River Gorge through the West Coast Range. This gorge is incised deeply into Ordovician conglomerate and Mt Read Volcanics, and crosses the major structural trend. The gorge appears to be an antecedent valley which is characteristic of active orogenic areas (Bloom, 1978). However, this area appears to have been orogenically inactive during the Quaternary and the gorge is probably a relic form developed during Tertiary uplift of the West Coast Range.

The major cirques formed on the eastern side of the West Coast Range occur on Mt Jukes, Mt Owen and Mt Geikie. None of the cirques are eroded deeply, and there is some suggestion that the altitudes of the cirque floors are geologically controlled (Bradley, 1954). A few smaller cirques also formed on the Eldon Range. Although they face both north and south those on the southern side of the range are considerably larger. The floors of these cirques appear to be at a similar altitude to the Late Palaeozoic-pre-Permian unconformity and may also be geologically controlled.

The major depositional landforms in the study area are moraines and outwash terraces in the valleys and screes on the slopes. The central part of the valley is dominated by middle Pleistocene outwash plains with gravels up to 60 m thick, and thick sequences of lacustrine sediments. In the lower part of the valley, near the Governor River, the large volumes of gravel have been repeatedly reworked by outwash streams of successive ice advances. The resulting stratigraphy and terrace geometry is complex and records outwash gravels from numerous ice advances.

Tasmania's position (40°– 44°S) between the Southern Ocean and the Tasman Sea means that it experiences the coldest climate in Australia. The humid maritime conditions are largely a consequence of the interception and orographic uplift of westerly maritime airstreams by the mountains. The West Coast Range area is the wettest region in temperate Australia, with a mean annual precipitation of 2000–3600 mm. The mean annual temperature at Queenstown is 10.9 °C. The mean monthly maximum of 22°C occurs in February, and the mean monthly minimum of 2.3°C occurs in July. There is no permanent snow or ice, and the present snowline lies well above the highest peaks of the West Coast Range.

The natural vegetation of the King Valley would be temperate rainforest dominated by *Nothofagus cunninghamii* with a component of *Lagarostrobos franklinii* near rivers (Kirkpatrick, 1977*a*). On the peaks of the West Coast Range the natural vegetation would be an alpine vegetation assemblage that includes the bolster plant *Donatia novae-zelandiae*, the coniferous shrubs, *Diselma archeri* and *Microcachrys tetragona*, the deciduous shrub *Nothofagus gunnii*, and graminoids. However, because of the combined effects of poor drainage, extensive burning of the vegetation during exploration, and selective logging activities in the area, the vegetation has been strongly altered (Kirkpatrick, 1977*b*, Kirkpatrick and Dickinson, 1984). Today the floor of the valley is mainly covered with epacridaceous heathland, *Gymnoschoenus sphaerocephalus* (buttongrass) sedgeland, and wet scrub of *Leptospermum scoparium*, *Melaleuca squamea* and *M. squarrosa*. The better drained slopes have regenerating *Eucalyptus simondsii*. There is very little *N. cunninghamii* present except on river banks where it occurs with remnants of *L. franklinii* and is partially protected from burning.

This study involves analysis of the glacial deposits in the middle and upper King Valley, a relatively small area (150 km^2) in the central West Coast Range where abundant exposures were produced during dam construction work by the Hydro-Electric Commission of Tasmania. The construction work presented the opportunity to study the stratigraphy of the deposits near the terminus of a major outlet glacier system for the purposes of testing, refining and possibly expanding the existing three glaciation model for Tasmania (Kiernan, 1983*a*; Colhoun, 1985*a*). The resulting work has produced one of the most detailed records of the effect of glaciation in a valley in the Southern Hemisphere (Fitzsimons, 1988; Fitzsimons and Colhoun, 1991).

The glacial legacy of Tasmania

The glacial legacy of Tasmania has been recognised since the late part of the last century (Dunn, 1894; Moore, 1894; David, 1926*a*, 1926*b*). Subsequent research by Lewis, based on the evidence of glacial erosion, suggested three periods of glaciation had occurred in Tasmania (Lewis, 1922, 1934, 1939, 1945). The three glaciation model developed by Lewis was succeeded by the work of Davies (1962), Derbyshire (1963), and Peterson (1968, 1969), who studied the depositional evidence for glaciation. These investigations concluded that most of the glacial deposits could be attributed to one glaciation (Gill, 1956; Jennings and Ahmad, 1957; Ahmad *et al.*, 1959; Davies, 1962; Peterson, 1968, 1969). Derbyshire *et al.* (1965) summarised much of the evidence in the Glacial Map of Tasmania. Although the map was based on the recognition of only one glaciation, it conceded that erratics beyond the limits of continuous drift were possible evidence for earlier ice advances (Derbyshire *et al.*, 1965; Derbyshire, 1968; Derbyshire and Peterson, 1971). In the Mersey–Forth area of northern Tasmania, Paterson (1965, 1966) and Paterson *et al.* (1967), recognised two glacial stages on the basis of the superposition of deposits, the degree of lithification, and the degree of chemical weathering. The older of the two glacial stages, the Lemonthyme Glaciation, is the oldest known glaciation in Tasmania and is probably of Tertiary age. After 1970 most research work has been concentrated on the central West Coast Range with the exceptions of recent work by Kiernan (1985) in the central highlands of Tasmania and by Hannan and Colhoun (1987) in northern Tasmania.

The glacial stratigraphy of the central West Coast Range of Tasmania has been summarised by Kiernan (1983*a*) and Colhoun (1985*a*). It consists of three glaciations, the Last or Margaret Glaciation which occurred during the Last Glacial Maximum (30 000–10 000 yr B.P.), the Henty Glaciation (>140 000 yr B.P.) and the Linda Glaciation (>730 000 yr B.P.) Subsequent fieldwork in the King Valley reported in this bulletin has shown that the stratigraphy is considerably more complex than was previously thought and that revision is necessary.

Mapping and classification of Quaternary glacial deposits

Although the study of glaciation in Tasmania has a long history (Banks *et al.*, 1987), the stratigraphic classification and mapping of glacial sequences has attracted relatively little attention. Most mapping has been achieved through a number of honours theses completed by students of the Department of Geography, University of Tasmania (Bowden, 1974; Sansom, 1978; Kiernan, 1980; Augustinus, 1982; Hammond, 1985; Benger, 1987). Though containing much useful material, they lack a consistent approach to mapping and stratigraphic classification. This, and the difficulty of obtaining enough exposure in the forested and mountainous terrain has made both the identification and correlation of glacial formations, even over very short distances of the order of tens of kilometres, difficult and in some cases speculative. Although none of the above studies resolved the complexity of the glacial stratigraphy, they have nevertheless provided the background against which the present findings can be compared. The much greater quantity and detail of stratigraphic information available from the King Valley now permits a more rigorous definition and classification of stratigraphic units.

The American Commission on Stratigraphic Nomenclature (1959) recommended the mapping of deposits using lithostratigraphic units that can be related to periods of climatic change. The lithostratigraphic units are defined in the hierarchy of formation, member, lentil, tongue and bed. In practice, few studies of glacial deposits subdivide members because sedimentary complexity and rapid facies changes make small scale divisions of little meaning for purposes of wider correlation. However, our experience has shown that the glacial deposits of western Tasmania can be mapped at the formation level as has been done in the 1:250 000 mapping programme of the New Zealand Geological Survey (Suggate, 1965*a*).

Although the ACSN recommendations regard morphologic units, 'aggradational surfaces rather than bodies of rock', as invalid stratigraphic units (*ibid.*, p. 665), many successful studies of glacial sequences have used morphologic units as primary mapping units. In New Zealand, moraines and outwash surfaces are frequently used as primary mapping units and their recognition may form the basis for defining a formation (Gage, 1958; Suggate, 1965*a*; Mabin, 1984).

More recently, The *International Stratigraphic Guide* (Hedberg, 1976), has recognised the Quaternary as a special field of stratigraphy and suggests that litho-stratigraphic classification based on indirect evidence such as geomorphology is a useful approach.

Initially, the study attempted to combine the stratigraphic and morphological methods and used morphostratigraphic units (moraines and outwash surfaces) as the primary mapping units. However, because of the erosion of some primary depositional landforms, several ice advances had to be reconstructed solely from sedimentary evidence (lithostratigraphic units). Biostratigraphic units have also been used and have been defined on the basis of distinctive vegetation assemblages determined by analysis of pollen and plant macrofossils. Empirical evidence indicates that a primary distinction can be made between non-forest pollen/vegetation assemblages during glacial stages and predominantly forest pollen/vegetation assemblages during interglacial stages.

The recent mapping of the King Valley permits a more rigorous approach to stratigraphic classification and warrants a formal definition of stratigraphic units which supplants and extends the existing designations. Because the sequence is discontinuous it was not possible to adhere strictly to the *International Stratigraphic Guide* (Hedberg, 1976). However, where possible the mapping and classification methods follow the guide in the definition of lithostratigraphic, morphostratigraphic and biostratigraphic units. Each formation has a type section and has been given a local geographic name (table 2). The physical characteristics of the glacial formations are summarised in Table 3. Climate-dependent chronostratigraphic stages have been based on the names of the glaciations and interglaciations. The definition of these stages follows the reasoning of Suggate (1965*a*, 1985*b*).

Table 2. FORMATIONS AND CLIMATIC STAGES OF THE KING VALLEY

Climatic Stage	*Formation*	*Interpretation*
Holocene	Long Marsh[2]	postglacial
Margaret Glacial Stage	Dante	ice advance
	Chamouni[2]	ice advance
Pieman Interglacial Stage[1]	Smelter[2]	interglacial
Henty Glacial Stage	Bull Rivulet[2]	ice advance
	David[2]	ice advance
	Nelson[2]	interstadial
	Cableway[2]	ice advance
Moore Glacial Stage[1]	Moore[2]	ice advance
	Pyramid[2]	ice advance
	Baxter[2]	interstadial
	Huxley[2]	ice advance
Regency Interglacial Stage[1]	Regency[2]	interglacial
Linda Glacial Stage	Thureau[2]	ice advance
Late Tertiary	Idaho[2]	preglacial

1. New climatic stage
2. New formation

Criteria for differentiating ice advances

Several criteria have been used to separate ice advances in the King Valley. The three main types of evidence are geomorphology, sedimentology, and dating. The most important of these are summarised in Table 3.

The altitude, and geographic distribution of moraines and outwash surfaces have been the most useful where they are preserved. However, because of the differential preservation of deposits, it has not been possible to use the same criteria to distinguish each formation. Several formations have been identified by their position within the stratigraphic sequence and by their sedimentary characteristics.

Comparison of the relative altitude of aggradation surfaces is based on the assumption that the highest aggradation surface at a locality was deposited during the maximum extent of a glacier advance (Suggate, 1965*a*). Identification of a series of outwash aggradation surfaces that can be traced upstream to either abrupt steepenings or terminations of the surfaces in moraine ridges can be taken to reveal the occurrence of multiple ice advances.

The geographic distribution of moraines can also assist interpretation of relative ages because more extensive ice advances destroy landforms of earlier, less extensive advances. A statistical analysis of the 'obliterative overlap' by Gibbons *et al.* (1984) suggests that up to seven out of ten end moraines may be destroyed in a succession of ice advances. This analysis serves as a warning that reliance on depositional morphology to determine the sequence of ice advances in a valley is likely to underestimate the number of advances. It is for this reason that reconstruction of depositional environments from fragmentary sedimentary evidence has proved to be both a necessary and important means of developing the glacial stratigraphy of the King Valley.

Several physical characteristics including particle size, pebble fabric and lithology were recorded to distinguish and interpret the origin of some deposits. Particle size gives the broad energy conditions of deposition. Pebble fabric analysis of glacigenic sediments has been used to give two different types of information which include direction of glacier movement, and the mode of sediment deposition (Boulton, 1971). While the use of till fabric for the reconstruction of ice movement direction has a long history of use (West and Donner, 1956), it is only recently that use of till fabric as a means of reconstructing the mode of deposition has become widespread. The growing body of evidence that suggests there is a strong relationship between mode of genesis and pebble fabric includes studies by Glen *et al.* (1957), Mills (1977), Lawson (1979), Haldorsen and Shaw (1982), Dowdeswell and Sharp (1986) and others. The transition from strong fabrics of melt-out and lodgement tills to the weaker fabric strengths of deformed and resedimented diamictons is interpreted as representing an increasing amount of disturbance or deformation subsequent to the release of sediments from ice (Lawson, 1979; Dowdeswell and Sharp, 1986). The effects of deformation on englacial fabrics summarised by Dowdeswell and Sharp are:

(1) a change from unimodal towards more random fabric patterns;

(2) a reduction in S_1 (see below) and concurrent decrease in S_3 eigenvalues;

(3) an increase in the frequency of steeply dipping clasts;

(4) an increase in the variability in fabric orientation and an increased deviation of the principal eigenvector from the observed ice flow direction.

Because the local direction of ice flow in the King Valley is constrained by topography and known from numerous striae, it is possible to use pebble fabric to reconstruct the mode of sedimentation of the deposits.

Usually two or more sets of 25 measurements were taken from an exposure, the number being based on Andrews and Smith's (1970) and Lawson's (1979) sampling schemes that provide a maximum 5° variance at the 95% confidence level. Sample sets consist of prolate shaped pebbles which are defined as having axial ratios of $a/b \geq 2$ and $b \approx c$. Measurements were plotted on equal area nets (lower hemisphere projection) and contoured using the method of Kamb (1959). The data were analysed using the eigenvector technique (Mark, 1973). The largest eigenvector V_1 indicates the direction of maximum clustering and the mean axis. V_3 indicates the direction of minimum clustering. V_2 is normal to V_1 and V_3. The significance values, S_1, S_2, and S_3, indicate the degree of clustering about the eigenvectors.

The provenance of tills was determined by counting the lithology of at least 100 pebbles between 15 and 60 mm in diameter from each sample. The pebble counts were grouped into major lithostratigraphic units that crop out in the study area and are plotted on pie diagrams. The majority of unidentified clasts belong to a group of Precambrian meta-sediments that have a wide variety of lithologies and are not well known.

Several dating methods including ^{14}C, amino acid racemisation and determination of the detrital remanent magnetisation of lake sediments were used to supplement and test the geomorphic and sedimentary evidence. The weathering rinds formed on widely distributed Jurassic dolerite clasts in the drifts were measured and used in an attempt to assess possible age differences since deposition of the glacial formations.

The value of ^{14}C dating in the development of the stratigraphy of the King Valley is limited because the ages of most glacial sediments lie beyond the usual limits of the technique around 40 000 yr. However, ^{14}C dating has been useful in defining the Pleistocene-Holocene boundary and the limits of late last glaciation ice advance.

Table 3. PHYSICAL CHARACTERISTICS OF THE GLACIAL AND RELATED FORMATIONS

Formation	*Location of type section*	*Distribution*	*Surface form*	*Lithology*	*Weathering rinds on Jurassic dolerite clasts in tills (mean and range)*	*Palaeo-magnetism*	*Age (yr B.P.)*
Thureau	Eastern edge of Thureau Hills	Remnants in the King Valley, large moraines in the Linda Valley	Eroded moraines or buried	Lacustrine sediments, ice-rafted tills, tills and outwash gravels. Composition indicates derivation from the Eldon Range	$\bar{x}$ = 44.1 mm 20 – 75 mm	Reversed	>730 000 ≈910 000?
Regency	2 km NW of the King–Governor confluence	Only known at one location	Buried	Drifted wood, leaves and peat	No dolerite	Not measured	>730 000
Huxley	Right bank of Baxter Rivulet	South of the Governor River	Small terrace remnant	Outwash gravel. Composition indicates derivation from the West Coast Range	No dolerite	Not measured	730 000 – 390 000
Baxter	Right bank of Baxter Rivulet	Only known at one location	Buried	Organic-rich sand.	No dolerite	Normal	730 000 – 390 000
Pyramid	Right bank of Baxter Rivulet	South of the Governor River	Small terrace remnants and ice-contact ridges	Outwash gravel and bouldery till. Composition indicates derivation from the West Coast Range	No dolerite	Not measured	730 000 – 390 000
Moore	200 m S of the King–Governor confluence	South of the Governor River	Buried	Coarse, ice-proximal, outwash gravel. Composition indicates derivation from the Eldon Range	$\bar{x}$ = 15.8 mm 10 – 25 mm	Not measured	730 000 – 390 000
Cableway	400 m N of the King–Governor confluence	South of the Thureau Hills	Eroded outwash surface and low hummocks	Outwash gravel, till and ice-contact stratified sediments. Derived from the Eldon Range	$\bar{x}$ = 6.9 mm 2.9 – 13 mm	Not measured	>140 000
Nelson	3 km N of the King–Nelson confluence	South of the Lyell Highway	Buried	Rhythmically laminated silt and mud	No dolerite	Normal	>140 000
David	3 km N of the King–Nelson confluence	South of the Lyell Highway	Large end moraine and outwash surface	Outwash gravel, lodgement till and bouldery till. Derived from the Eldon Range	$\bar{x}$ = 11.2 mm 5 – 16 mm	Normal	>140 000
Bull Rivulet	5 km NE of Mt Owen	Small area 2 km south of the Lyell Highway	Small moraine and outwash surface	Bouldery till. Derived from the Eldon Range	Not measured	Not measured	>140 000
Smelter	3 km E of Mt Jukes	Only known at one location	Buried	Drifted leaves and organic silt	No dolerite	Not measured	>140 000
Chamouni	2 km S of Dante–King confluence	Large terraces north of Lyell Highway	Outwash terraces	Laminated silts, outwash gravel and lodgment till. Derived from the Tyndall Plateau	$\bar{x}$ = 1.5 mm 0.9 – 2.8 mm	Not measured	>48 000
Dante	1 km NE of Dante–King confluence	Limited to the Comstock Valley	Small outwash fan	Outwash gravel, sand and till. Derived from the Tyndall Plateau	$\bar{x}$ = 1.5 mm 0.9 – 2.7 mm	Not measured	<18 800

The important ^{14}C dates in the King Valley and surrounding area are summarised in Table 4.

Since it was anticipated from the stratigraphic position of many wood samples that they would be well beyond the limits of ^{14}C dating, several were dated to prove that they really were beyond the ^{14}C age range. Most of these assays produced finite dates from 30 000 to 39 000 yr B.P. (table 4) even though the counter at The N. W. G. Macintosh Centre for Quaternary Dating at Sydney University has a nominal range of 50 000 yr. The wood samples were almost certainly contaminated by modern humic acids which are mobile in humid environments such as the West Coast of Tasmania. This and other problems of contamination of datable materials in Tasmania have been outlined by Colhoun (1986). Because of these difficulties all ^{14}C over 30 000 yr B.P. in the King Valley must be regarded as infinite.

An attempt was made to date wood samples that were significantly older that the range of radiocarbon dating by the amino acid method. Amino acid dating of wood

Table 4. RADIOCARBON DATES IN THE STUDY AREA

Date (yr BP)	Lab. Code	Location	Source	Comments
26 480 ± 800	W 323	Linda Creek [CP835422]	Gill (1956)	Wood from rhythmites
>40 000	NZ 348	Linda Creek [CP835422]	Grant-Taylor and Rafter (1963)	Wood from alluvial sediments
>40 000	R 488	Linda Creek [CP835422]	Banks *et. al.* (1977)	Wood from alluvial sediments
>48 500	ANU 3413	Linda Creek [CP835422]	Colhoun (1985b)	Wood from alluvial sediments
27 800 ± 700	ANU 2480A	Linda Creek [CP835422]	Colhoun (1985*b*)	Wood from alluvial sediments
23 100 ± 600	ANU 2480B	Linda Creek [CP835422]	Colhoun (1985*b*)	α-cellulose of ANU 2480A
9 050 ± 120	SUA 1358	Tyndall Plateau	Macphail (1986)	Minimum age of final high-altitude deglaciation
18 800 ± 500	ANU 2533	Dante Rivulet [CP902459]	Kiernan (1980)	Last glacial maximum
21 180 ± 370	SUA 2154	Dante Rivulet [CP902459]	Kiernan (1985)	Last glacial maximum
20 100 ± 470	SUA 2155	Dante Rivulet [CP902459]	Kiernan (1985)	Last glacial maximum
19 000 ± 170	SUA 2856	Dante Rivulet [CP902459]	Fitzsimons and Colhoun (1991)	Last glacial maximum
37 800 +800/−700	SUA 2469	King Valley [CP872389]	Colhoun and van de Geer (1987)	
32 800 +400/−700	SUA 2392	King Valley [CP870307]	This Bulletin	Minimum age of David Formation outwash
39 300 +800/−700	SUA 2393	King Valley [CP875306]	This Bulletin	Minimum age of David Formation outwash
2150 ± 90	SUA 2487	Nelson River [CP917354]	This Bulletin	Recent river erosion
35 200 +800/−700	SUA 2488	King Valley [CP878305]	This Bulletin	Minimum age of David Formation outwash
20 660 ± 280	SUA 2521	Newall Creek [CP800314]	This Bulletin	Peak of late last glaciation
2520 ± 80	SUA 2597	King Valley [CP881323]	This Bulletin	Holocene slope deposits
48 700 +2900/−2100	SUA 2599	King Valley [CP886423]	This Bulletin	Minimum age of Chamouni Formation
>21 000	SUA 2621	Newall Creek [CP800314]	This Bulletin	Basal date of organic sediments
12 250 ± 90	SUA 2415	King Valley [CP892378]	This Bulletin	Base of Holocene gravel
13 010 ± 130	SUA 2723	King Valley [CP880307]	This Bulletin	End of the Last Glaciation

involved the determination of the dextro to laevo stereoisomers (D/L ratio) for aspartic acid. The dates reported are part of an experimental study by Dr Brad Pillans of the Department of Geology, Victoria University of Wellington. The method used is reported by Pillans (1983) and the results are regarded as minimum values that are stated broadly in terms of isotope stages.

The measurement of detrital remanent magnetisation gives the polarity of the earth at the time of sediment deposition. To convert this information to age it must be compared with the known conditions of the earth's magnetic field that has been dated by some other method. The results reported in this bulletin are for measurements made on laminated silts and muds deposited in glacial lakes, although diamictons such as till and sediment flows may also carry a stable remanence (Easterbrook, 1988). Specimens for palaeomagnetic analysis were collected in 8 cm^3 plastic cubes pressed or lightly tapped into the sediment surface. Twenty to thirty specimens were collected from most sites. These specimens were taken in groups or spreads across the sites and were arranged to ensure maximum representation of the vertical and lateral extent of the deposits. Areas of obvious disturbance resulting from slumping and ice pushing were avoided, as were areas affected by chemical weathering.

Wherever possible, only flat-lying sediments were sampled; tilted sediments were only sampled when there were no other suitable sites. All specimen orientations were taken with a magnetic compass as it was not feasible to use a sun compass. The Nelson Formation was sampled from drill cores obtained by the Department of Main Roads, as well as from exploration trenches.

All magnetic measurements were made at the Black Mountain Palaeomagnetic Laboratory of the Research School of Earth Sciences, Australian National University in Canberra, under the direction of Dr C. Barton of the Bureau of Mineral Resources.

Natural remanent magnetism (NRM) and susceptibility were measured for all specimens. Representative samples were demagnetised over a range of values. NRM, a summation of all components of specimen remanence acquired by natural processes (Tarling, 1983), was measured using a superconducting magnetometer.

Extremely strongly magnetised specimens were measured using a spinner magnetometer. Magnetic susceptibility was measured using a susceptibility bridge. An alternating field demagnetiser was used to demagnetise representative samples systematically. This step-wise demagnetisation was used to determine the stability of magnetisation of specimens and the values at which the individual site specimens were to be demagnetised.

NRM and susceptibility values were used to calculate the modified Konigsberger ratio for samples. This ratio provides an indication of the sample's ability to maintain a stable remanence. Values less than 0.1 indicate a poor capability whilst values greater than one indicate a good capability (Collinson, 1983). The importance of palaeomagnetism in defining the King Valley stratigraphy is twofold. Firstly, it has confirmed that some deposits beyond the range of ^{14}C dating are normally magnetised, and are therefore believed to have been deposited during the Brunhes Chron (0–730 ka B.P.; Bowen, 1978). Secondly, deposits of the Thureau Formation are consistently magnetically reversed and are believed to have been deposited during the Matuyama Chron, i.e. before 730 000 yr B.P. This boundary is used to separate deposits of Middle and Early Pleistocene age (Mankinen and Dalrymple, 1979).

Weathering rinds on Jurassic dolerite were measured to extend earlier work by Kiernan and to test the utility of the method on a relative stratigraphy determined by other methods. Fifty measurements were made at each site in accordance with the methods used by Colman and Pierce (1981) and Kiernan (1980, 1983*a*, 1985). Water absorption and specific gravity of clasts was also measured using the same methods as McGregor (1981) and Augustinus (1982), but these techniques were abandoned at an early stage because they provided similar information to the weathering rind data.

Pollen analysis of organic sediments found in the study area proved to be a valuable means of separating ice advances and glacial deposits from non-glacial interstadial and interglacial deposits. The pollen samples were prepared by the Faegri and Iversen (1975) method. To enable both percentage and concentration calculations to be made, exotic *Lycopodium clavatum* spores were added to the samples (Stockmarr, 1971). A minimum pollen sum of 300 grains included all terrestrial woody and herbaceous taxa plus tree ferns. The taxonomic nomenclature follows Curtis (1963, 1967), Curtis and Morris (1975), Willis (1970) and Wakefield (1975).

The glacial system of the central West Coast Range

The extents attained by glaciers during successive ice advances in the King Valley are shown on Figure 2. Ice flowed into the King Valley from numerous cirques on the Eldon Range and West Coast Range, and from a thin ice cap on the Tyndall–Lake Dora Plateau. During the more extensive advances the King Glacier split into four distributary lobes and flowed south down the King Valley, east and south into the Nelson Valley and into the Collingwood catchment, west up the Comstock Valley, and west up the Linda Valley.

During the greatest ice advances of the Linda Glaciation ice from the main lobe flowed down the King Valley and terminated on the Crotty Plain near the entrance to the King River Gorge in the West Coast Range (fig. 2). The absence of erratic Jurassic dolerite and Permian sediments in numerous exposures between the Governor River and the Andrew Divide, suggests that ice from the Tyndall Plateau flowed no further south than the limit drawn on Figure 1. There is no evidence to suggest ice from the King Valley crossed the Andrew Divide.

Ice that flowed eastward across the Little Eldon Range appears to have crossed into the Collingwood catchment where it was confluent with ice that flowed from the headwaters of the Franklin catchment (Kiernan, 1985).

During the less extensive ice advances of the Henty Glaciation a distinct southward flowing lobe, parallel to the main lobe of the King Glacier, formed in the Nelson Valley (fig. 2).

The Comstock lobe, together with ice from the Tyndall Plateau, flowed some distance down the Queen Valley at least as far as Lynchford. Kiernan (1980) suggested that ice from the Linda lobe flowed into the Queen Valley but the moraines at the head of the Linda Valley suggest that it did not breach Karlsons Gap. This is confirmed by the lack of Jurassic dolerite erratics and the steep, unglaciated valley of Conglomerate Creek west of the gap. The largest distributary lobe of the central West Coast Range glacial system flowed down the King Valley and is the main concern of this bulletin.

Several small glaciers formed on the West Coast Range and flowed down into the King Valley. Ice from Mt Owen flowed into the Tofft Valley between the Thureau Hills and the West Coast Range where it coalesced with the King Glacier during its more extensive advances. Ice from Mt Jukes, which carried the largest of the cirque glaciers on the southern part of the West Coast Range, formed two piedmont lobes separated by a rock ridge (fig. 2). The southern lobe, which flowed in the Fish Creek catchment, appears to have been the most extensive, though it is unlikely that it extended into the Governor and Andrew river valleys.

The multiple ice sources that contributed to the glacial system in the King Valley makes the stratigraphic classification of the deposits complex. The small glaciers on Mt Jukes and Mt Owen were much smaller than the distant ice which formed on the Tyndall–Lake Dora Plateau and Eldon Range areas that fed the King Glacier. The proximity of the local cirque glaciers to the middle King Valley meant there was a time lag between the deposition of sediment fluxes associated with ice advances from the cirques and from the main valley during the same phase of glacier advance. Andrews (1975) estimated the time lag for a small temperate glacier to be of the order of 10^2 years compared to 10^3 years for an ice mass the size of the Greenland ice sheet. The same principle applies to the glacial system of the King Valley though the sizes of the glaciers of the West Coast Range of Tasmania were much smaller. Although the volume of the glaciers cannot be accurately assessed, the King Glacier is estimated to have been 25 times larger in area than the Jukes Glacier. Because of the effects of time-lag, the lithostratigraphic units from different sources, but related to the same major climatic event, can overlie or lie adjacent to each other. This phenomenon is recorded by a series of sections at Baxter Rivulet where the Pyramid and Moore Formations appear to have been deposited during the same phase of glacial advance by ice from different sources.

THE GLACIAL AND INTERGLACIAL SEQUENCE OF THE KING VALLEY

Nine ice advances from four glaciations have been identified in the King Valley (table 2). The extent and distribution of deposits associated with each advance is summarised in 1:25 000 geological maps (Corbett *et al.*, 1989). The relationships between the deposits formed by the distributary lobes of the King Glacier and those formed by glaciers from Mt Jukes and Mt Owen are summarised by Table 5. The locations of the sections described are given as AMG references which may be found on the Tasmap 1:100 000 Franklin sheet (8013).

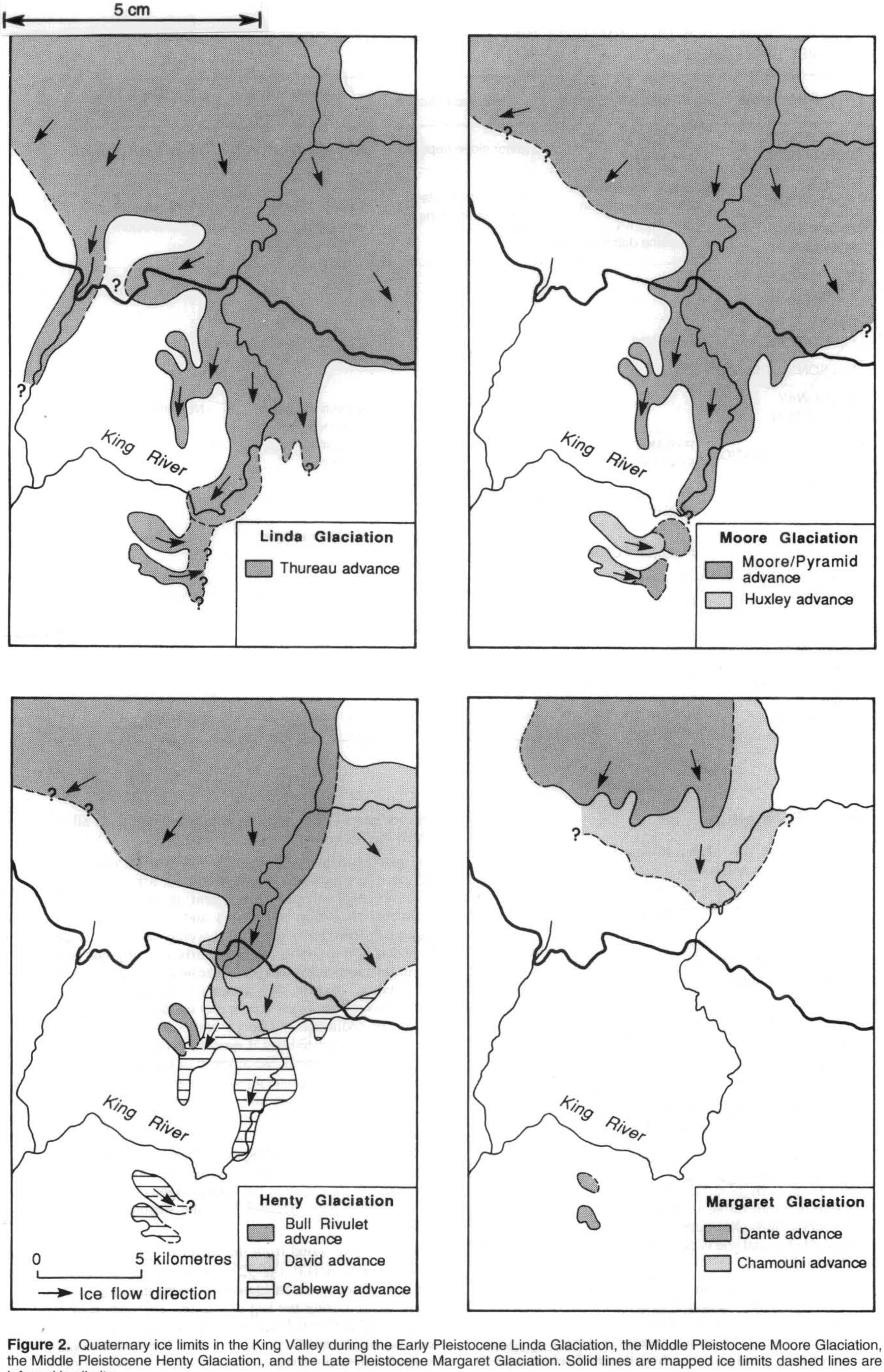


Figure 2. Quaternary ice limits in the King Valley during the Early Pleistocene Linda Glaciation, the Middle Pleistocene Moore Glaciation, the Middle Pleistocene Henty Glaciation, and the Late Pleistocene Margaret Glaciation. Solid lines are mapped ice limits dashed lines are inferred ice limits.

Table 5. CORRELATION OF FORMATIONS WITH DEPOSITS OF THE CIRQUE GLACIERS AND DISTRIBUTARY LOBES

<table>
<tr><th>King Glacier</th><th>Mt Jukes Glacier</th><th>Mt Owen Glacier</th><th>Nelson River distributary</th><th>Linda Valley distributary</th></tr>
<tr><td>LONG MARSH FORMATION</td><td>Minor stream bed instability</td><td>Minor slope deposits</td><td>Minor stream erosion</td><td>Minor slope deposits</td></tr>
<tr><td>DANTE FORMATION</td><td rowspan="2">Small outwash fans
Small moraines in cirques
Moraine dammed lakes</td><td rowspan="2">Small outwash fans
Terminal moraine in cirque</td><td rowspan="2">Erosion of older sediments</td><td rowspan="2">Slope deposits and minor erosion</td></tr>
<tr><td>CHAMOUNI FORMATION</td></tr>
<tr><td>BULL RIVULET FORMATION</td><td rowspan="4">Not found</td><td rowspan="7">Not found</td><td rowspan="5">Numerous end moraines and extensive glacial deposits</td><td rowspan="7">Not differentiated</td></tr>
<tr><td>DAVID FORMATION</td></tr>
<tr><td>NELSON FORMATION</td></tr>
<tr><td>CABLEWAY FORMATION</td></tr>
<tr><td>MOORE FORMATION</td><td>PYRAMID FORMATION</td></tr>
<tr><td>Not found</td><td>BAXTER INTERSTADIAL</td><td rowspan="2">Not differentiated</td></tr>
<tr><td>Not found</td><td>HUXLEY FORMATION</td></tr>
<tr><td>REGENCY FORMATION</td><td>Not found</td><td>Not found</td><td>Not found</td><td>Organic remains at Lynchford</td></tr>
<tr><td>THUREAU FORMATION</td><td>Extensive terraces in Governor and Andrew Rivers</td><td>Not found</td><td>Extensive, highly weathered ice-contact deposits</td><td>Large moraines in the head of the Linda and Queen Valleys</td></tr>
<tr><td>IDAHO FORMATION</td><td>Not found</td><td>Not found</td><td>Not found</td><td>Not found</td></tr>
</table>

Idaho Formation

Sediments of the Idaho Formation are the oldest unconsolidated sediments known in the study area. Although their age is not accurately known, they ante-date sediments of the Linda Glaciation and are probably of Pliocene age.

The type section of the Idaho Formation is one kilometre north of the town of Gormanston at CP835422, in the valley south of Idaho Creek. The section shows a series of locally derived fluvial sands and gravels overlying a palaeosol with *in situ* tree stumps. The sediments are exposed on an eroded spur in a gully and consist of 12 m of bedded fluvial gravel and sands that dip at ≈23°SW towards a rock wall (fig. 3). Although the contact is not seen, the sediments are inferred to lie beneath highly weathered Thureau Formation glacigenic deposits which crop out above and south-east of the section.

The lower part of the section consists of coarse gravelly sand overlain by a 1.5 m thick palaeosol with stumps of trees in growth position. Pollen from the palaeosol consists of a wide range of forest taxa, some of which have a Tertiary affinity (Kiernan, 1980). The palaeosol contains abundant charcoal and is developed on coarse quartz sand and granules.

The palaeosol is overlain by 400 mm of coarse sand that contains numerous wood fragments, up to 15 mm in diameter, that are derived from the palaeosol. The remainder of the sediments consist of layers of coarse massive sand interbedded with well-sorted pebble gravel and occasional thin layers of laminated sandy silt all of which dip between 15° and 23°SW (fig. 3).

The quartz and quartzite lithology of the fluvial sediments indicates they are locally derived (fig. 3), which contrasts with the high erratic component in the glacigenic sediments that crop out above and south-east of the section. Pebbles in the gravels are well rounded while the granules and pebbles in the sands are angular and subangular indicating that they have not travelled far. The provenance, texture and geometry of the sediments distinguish them from the overlying and surrounding glacigenic sediments. Although earlier interpretations of this section suggested that it was at least partly of glacial origin (Kiernan, 1980), the absence of erratic rock types shows they were deposited prior to the overlying glacial sediments. The geometry and stratigraphic position of the sediments below the Thureau Formation sediments indicates that the Idaho Formation sediments are an eroded, non-glacial, probably Late Tertiary, inlier within Thureau Formation glacial lake sediments. The sequence probably accumulated on the foreset slope of a delta and the sediments were derived from the surrounding rock slopes.

Initial dating of wood from the palaeosol gave a date of 26 480 ± 800 yr B.P. (W-323), (Gill, 1956). For many years this date was considered to record the maximum extent of ice during the last glaciation. Two subsequent dates of >40 000 yr B.P. (R-488 and NZ-348), (M. Banks pers. comm.; Grant-Taylor and Rafter, 1963), contradict the earlier date and suggested problems in interpretation.

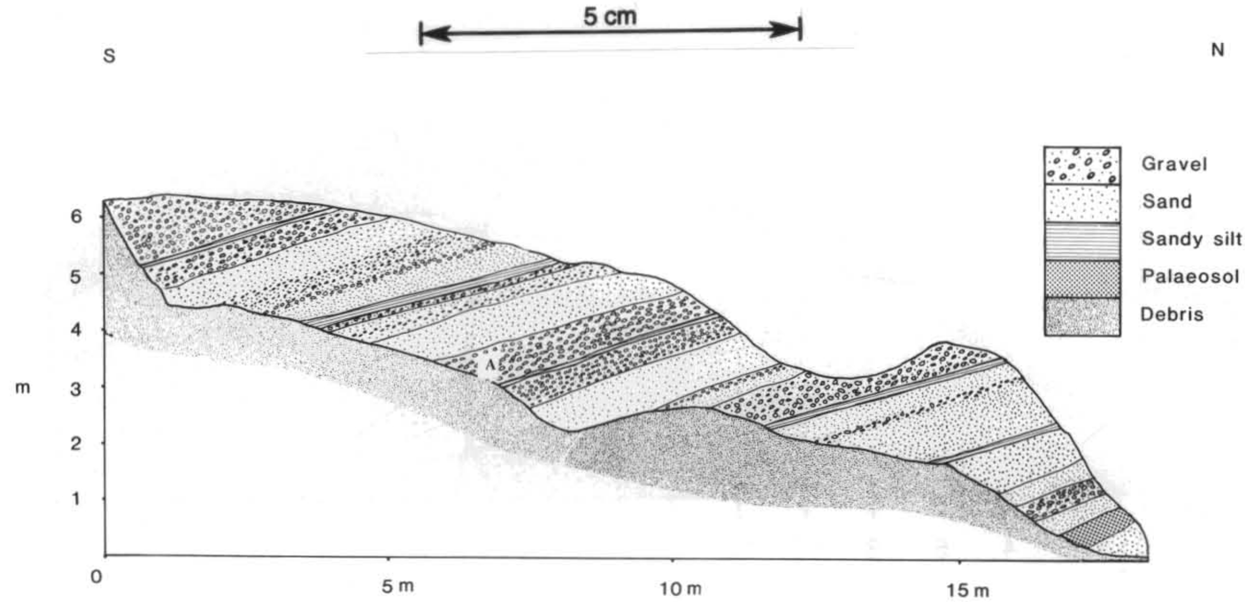


Figure 3. Interbedded sands, gravels and silts of the Idaho Formation. The letter A corresponds to fabric analyses and samples taken for pebble counts.

Recent dating of whole wood at 27 800 ± 700 yr B.P. and the alpha cellulose fraction at 23 100 ± 600 yr B.P. suggested a serious contamination problem (Colhoun, 1985*a*). Determination of the $\delta^{13}C/\delta^{12}C$ values of the dated wood suggests that all ^{14}C dates are suspect (*ibid.*, p. 54). Although the dating problem is not completely resolved a further date of >48 500 yr B.P. (ANU 3413) suggests that the wood is probably of infinite radiocarbon age. Amino acid dating of the wood was attempted but the samples were too degraded to give a reliable date (B. J. Pillans, pers. comm., 1987).

Thureau Formation

During the Thureau advance, ice reached its maximum extent in the King Valley and split into four distributary lobes. (fig. 2). At this time ice extended 19 km down the King Valley past the Thureau Hills to the entrance of the King River Gorge through the West Coast Range, where it terminated near or against ice from Mt Jukes. The King River Gorge acted as the major meltwater channel outlet and carried large amounts of sand and gravel that accumulated near Newall Creek, on the western side of the West Coast Range.

Thureau Formation sediments are preserved in five principal areas, the middle King Valley near the Thureau Hills, the Linda Valley, the Queen Valley, the lower King Valley, and the Nelson Valley. The most extensive sediments accumulated in ice-contact lakes in the Linda and Nelson Valleys which were outside the limits of subsequent ice advances. Most other exposures have been subject to erosion by younger ice advances and are mostly buried by younger outwash gravels, fluvial deposits and slope deposits.

The type section of the Thureau Formation is the northern-most road cutting on the Crotty Road in a series of exposures of highly weathered till buried by slope deposits from the adjacent Thureau Hills. The exposure at CP882353 is cut through a steep colluvial fan surface that buries a suite of glacial sediments. The underlying glacial sediments consist of massive till overlain by a series of flow tills, massive sands and laminated lake sediments (fig. 4). The whole sequence was penecontemporaneously deformed by the melting of ice on which it was deposited.

The diamicton at the base of the section is coarse, massive, matrix-supported and contains small lenses of stratified sands and gravels. The pebble fabric of the diamicton is widely dispersed, indicating post depositional deformation or redeposition by mass movement processes (fig. 5A and 5B).The overlying 1.2 m-thick diamicton is unsorted, has occasional flow noses, and has rudimentary flow banding that dips north at 15°. Jurassic dolerite clasts, which form a significant part of the lithology of this deposit, are highly weathered (fig. 6) and have weathering rinds with a mean thickness of 54.5 mm and standard deviation of 28.2 mm. The matrix is also highly weathered and bleached to a pale grey colour. The pebble fabric has a weak concentration around 267° that is apparently unrelated to the glacier flow direction (≈180°) (fig. 5C). The contact with the overlying sediment flow is sharp, and is marked by a scoured surface that is parallel to the flow banding and bedding of both diamictons.

The dipping diamicton is overlain by another diamicton that is to 1.3 m thick and consists of 4 beds of fine clast-supported pebbles with intervening layers of highly weathered and bleached, pebbly diamictons that appear to be sediment flows (fig. 4, 6A). The pebble fabric is dispersed, and appears to reflect the palaeoslope on which the sediment flows and meltwater deposits formed (fig. 5D). The provenance of the clasts in the sediment flows is dominated by West Coast Range and local rocks with rare pebbles of erratic Jurassic dolerite and Permian sediments. The contact with the overlying sand is sharp and the diamicton appears to bury a depositional surface that dips north at 10 to 16°.

The three diamictons are overlain by up to 3.5 m of moderately well-sorted coarse white quartz sand. The top 1.5 m is horizontally laminated with thin beds of silty sand and the remainder is massive. A small exposure in a drain 10 m north of the section exposes 3 m of laminated fine sand and silty sand. Although the relationship between the two exposures is not clear these sands are interpreted as a northward thickening wedge of the sand in the main section. This interpretation is crucial to understanding the temporal relationships of the sections because the sands have a normal detrital remanent magnetisation while the overlying silts have a reversed polarity. The contact between the sand and the overlying diamicton is sharp and eroded.

The sand is overlain by a diamicton up to one metre thick that consists of a thin bed of rounded pebbles in a matrix of grey silt similar to the overlying laminated silts. It dips north at 11°. It is structureless except for the occasional fault which extends from the overlying silt (fig. 4). The

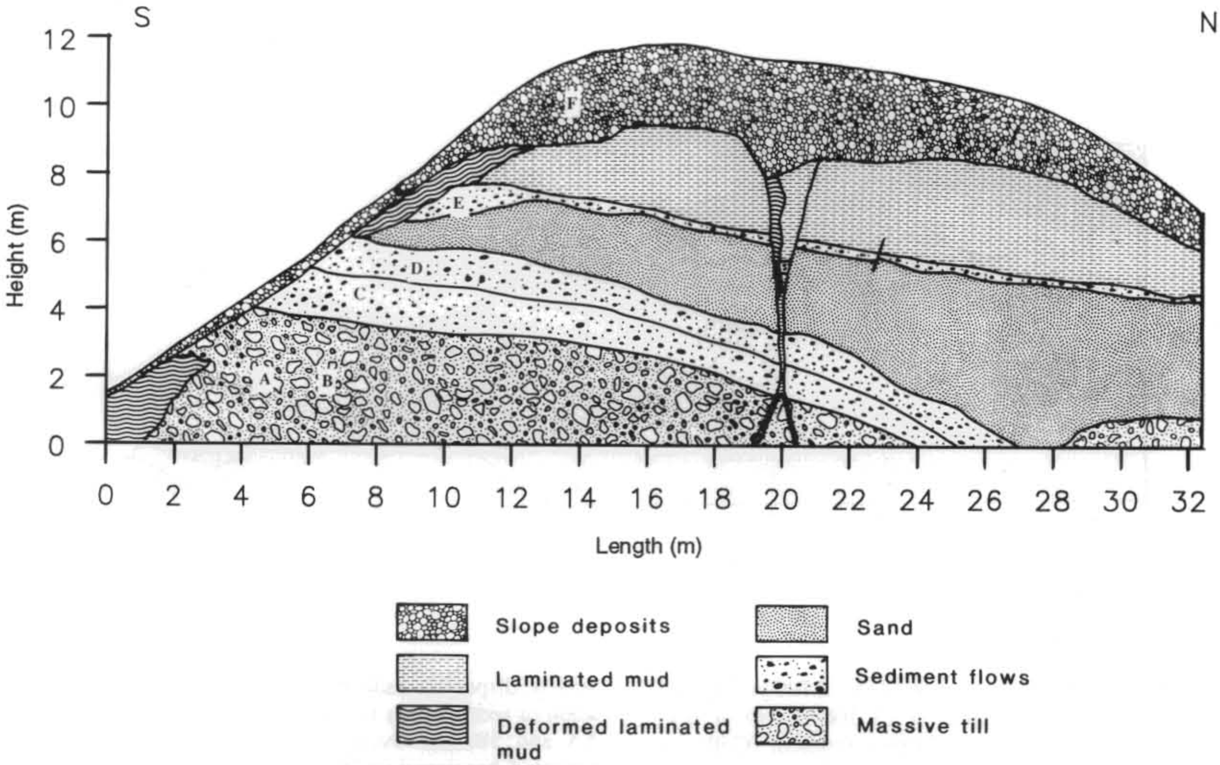


Figure 4. Type section of the Thureau Formation. A massive diamicton overlain by multiple sediment flows, massive and laminated sand, a sediment flow, laminated silts and slope deposits. Letters refer to location of fabric analyses.

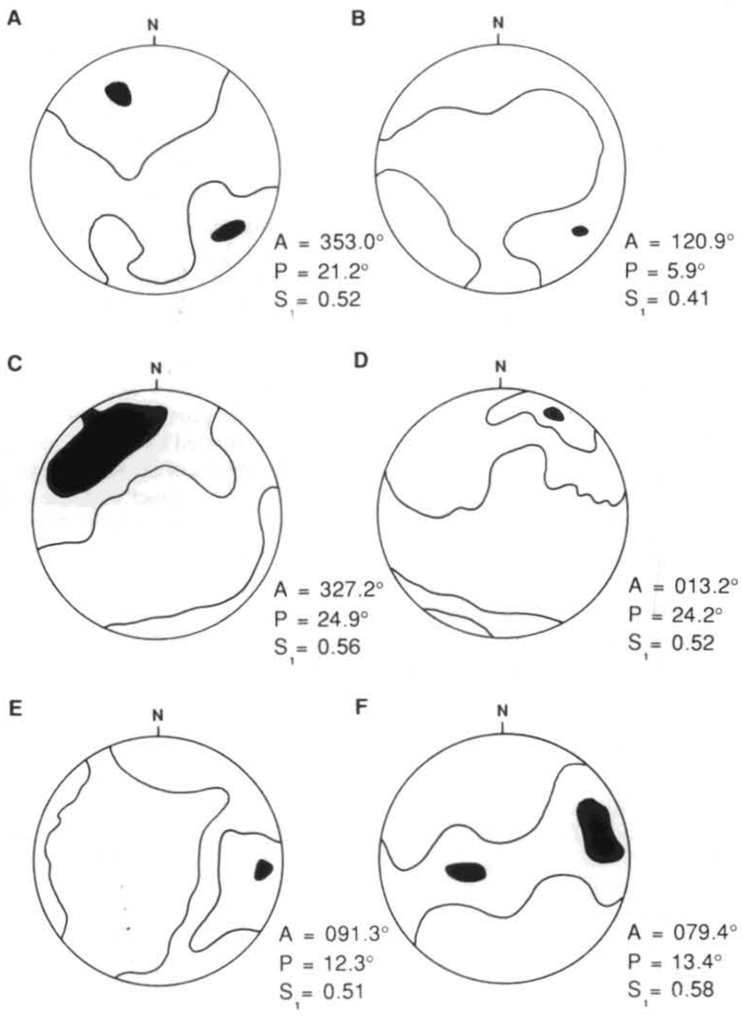


Figure 5. Contoured equal area plots of pebble fabrics from the type section of the Thureau Formation. See Figure 4 for sample locations. Contour interval is 2 standard deviations and the zone of maximum clustering is black.

5 cm

A

B

C

Figure 6. The Thureau Formation type section. A, Interbedded sediment flows and sheet flow deposits. B, Highly weathered Jurassic dolerite cobbles below a sediment flow. C, Wedge-shaped clastic dyke in Thureau Formation till.

5 cm

lithology of the diamicton consists largely of rocks with a local and West Coast Range provenance and rare erratic clasts. The pebble fabric is weak, unrelated to the ice flow direction which was from north to south (fig. 5E), and appears to reflect the slope it was deposited on. Numerous beds of the underlying sand are truncated by the diamicton, which may be a thin sediment flow or ice-rafted diamicton deposited during the onset of deposition in the lake.

The thin diamicton is overlain by up to 3 m of dark green and dark grey and green laminated silts. The dark green laminae are silts and the dark grey laminae are clayey silts according to the classification of Folk *et al.* (1970), (fig. 7). Individual laminae are well sorted and there is upward-fining textural differentiation in the green laminae.

Small-scale primary and secondary structures are common in the silts and include convolute lamination and reverse faults. The syndepositional nature of the convolute lamination is apparent from its occurrence in specific beds which are overlain by undisturbed beds. The trend of numerous high-angle reverse faults in the silts are strongly clustered toward 279° and dip between 70 and 80°. These faults were probably associated with other deformation structures observed in the section. On the southern side of the section a normal fault with a throw of 6.5 m is inferred from the position of intensely folded laminated silts near the bottom of the section (fig. 4). A wedge-shaped block failure that is down faulted by 1.1 m and intrudes the underlying diamicton and sands also suggests a period of deformation (fig. 4, 6C). The structure bifurcates at 10 m depth and its margins are stained by humic acid and iron.

Twenty specimens were taken from the silts for palaeomagnetic analysis. Most of specimens were step-wise demagnetised and the remaining specimens were demagnetised at 20 mT. All samples were weakly magnetised with natural remanent magnetism (NRM) values ranging from 0.15 to 1.22 μG. Slightly more than half the specimens had a fair capability of maintaining a stable remanence, whilst the remainder had only a poor capability. All specimens had a reversed polarity which suggests deposition prior to 730 000 yr B.P. (fig. 8A).

Twenty-five specimens were also taken from the laminated fine sands and silty sands exposed in a small creek 10 m north of the Thureau Formation type section. These sediments have been interpreted as a northward thickening of the sand from the type section. Most specimens from this site were also weakly magnetised, with NRM values ranging from 0.07 to 3.1 μG. The majority of specimens had a fair capability of maintaining a stable remanence. Half the specimens were stepwise demagnetised whilst the remaining specimens were cleaned at 35 mT. Almost all the specimens had a normal polarity (fig. 8B).

If these sediments represent the laminated mud of the type section (fig. 4), then they would have been deposited during one of the normal events of the Matuyama Chron. It is difficult to escape the interpretation that the section crosses a palaeomagnetic boundary, possibly from the Jaramillo or Olduvai normal events into the reversed part of the Matuyama Chron. The Jaramillo event is known to have occurred between 950 000 and 890 000 yr B.P. and the Olduvai event between 1.83 and 1.62 Ma B.P. (Bowen, 1978).

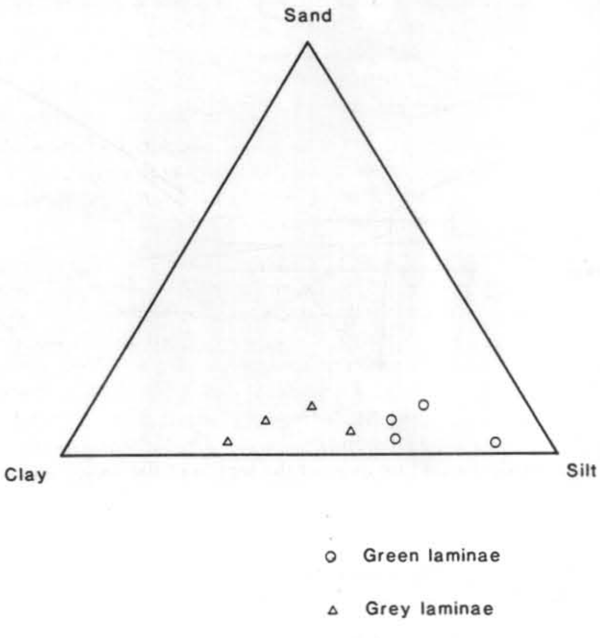


Figure 7. Ternary diagram of the particle size of laminated silts of the Thureau Formation type section.

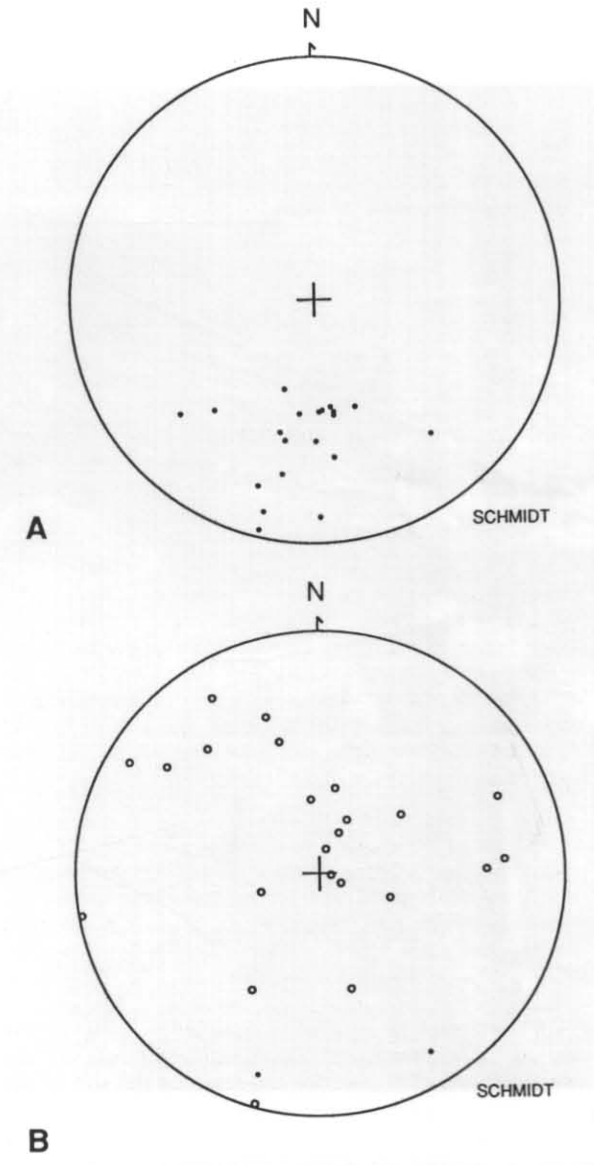


Figure 8. A, Stereoplot of the detrital remanent magnetisation of laminated muds from the type section of the Thureau Formation. B, Stereoplot of the detrital remanent magnetisation of laminated sediments 100 m north of the Thureau Formation type section.
Open circles indicate normal polarity, solid circles indicate reversed polarity.

5 cm

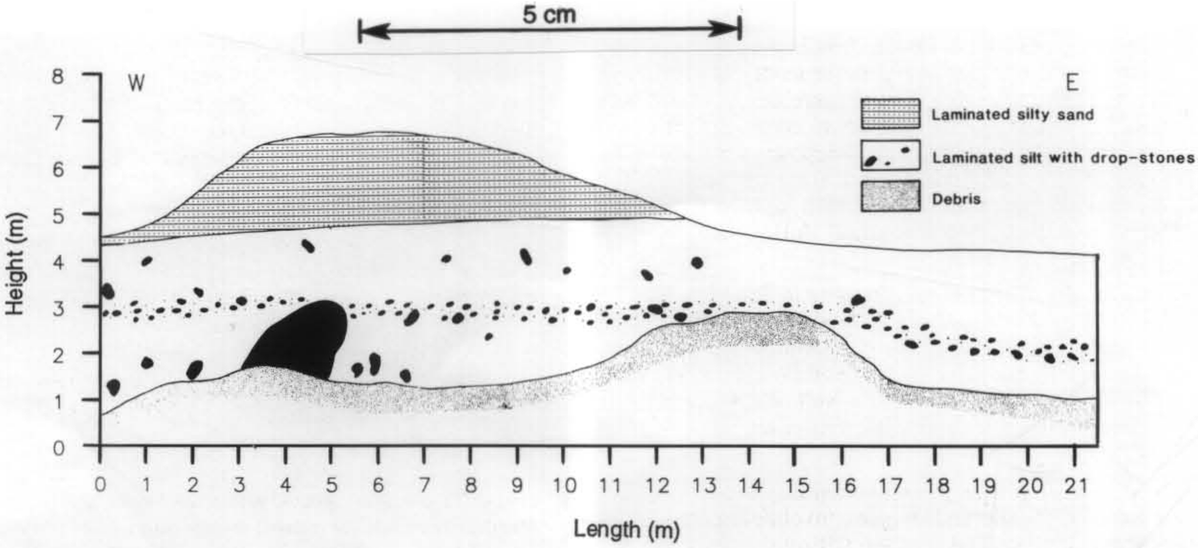


Figure 9. Thureau Formation sediments exposed at Gormanston Football Field. Laminated sandy silts overlie laminated mud containing sediment flows and ice rafted diamictons which rest on massive till.

The silts of the type section are overlain unconformably by poorly-sorted slope deposits derived from the Thureau Hills (fig. 4). Although the contact with the underlying lake silts is certainly erosional, there is no indication of the period of time that separates their deposition. The pebble fabric of the gravels shows a bimodal girdle pattern (fig. 5F). The gravel is coarse and deficient in fines. This reflects the siliceous source, the short distance, and high energy environment of transportation.

The Thureau Formation type section records an ice-contact, deglacial environment. The sediments were probably deposited in a supraglacial position on melting ice. The reconstructed sequence of events in the deposition of the sediments is as follows. The terminus of the glacier became almost stagnant and deposition of a massive sediment flow or deformation of a melt-out till occurred. Then deposition of multiple sediment flows and sheet flows occurred within a depression that became filled with water which increased in depth as the laminated sands accumulated. Subaqueous debris flows occurred within the developing kettle hole and were followed by the accumulation of laminated silts which settled out from small turbidity currents. The erosion and burial of the glacial sediments by slope deposits occurred during and after the deformation of the entire section when the buried glacier ice melted. The change from subaerial sediment flows to subaqueous deposition may represent sedimentation in a deltaic environment as the depth of water gradually increased due to the melting of ice under the accumulating sediment.

Deformation of the sediments is very similar to Sandford's (1959) type 1 model experiments which have been adapted for explanation of ice-contact deformation by McDonald and Shilts (1975). The model shows the effect of gentle upwarping, which is the equivalent of removal of support at the ends of the section by melting. The melting caused the formation of a wedge-shaped graben. The central wedge failure in the Thureau Hills (fig. 6C) section is almost identical. Faults in the chaotically deformed laminated sediments on the southern end of the section suggest that the size of the buried ice blocks must have been large to have caused a throw of 6.5 m. However, such deformation is very localised and large parts of the section are relatively undeformed.

Extensive highly weathered Thureau Formation sediments and landforms occur in the upper part of the Linda Valley. Most of the sediments consist of chaotically deformed ice-rafted diamictons and sediment flows interbedded with laminated lake silts.

The section at Gormanston Football Field [CP836412] consists of dark grey laminated lake silts and subaqueous sediment flows that rest on highly weathered massive till (fig. 9–10). At the base of the section, highly weathered Jurassic dolerite boulders up to 1.8 m in diameter rest in a silty matrix. The diamicton overlies angular Cambrian volcanic rock detritus which may have resulted from ice fracturing. The massive diamicton is overlain by 1.5 m of grey laminated silts with numerous dropstones (fig. 9). Within the silts are discrete beds of pebbles that appear to record episodes of ice-rafting.

On the eastern side of the section thick gravel lenses occur in the silt. These isolated lenses are associated with intense deformation of the silt which is folded, faulted and brecciated. The intense marginal deformation around the lenses suggest they formed as subaqueous sediment flows.

Six specimens were taken from sandy silts in the Gormanston Moraine for comparison with the samples from the Thureau Formation type section and with those taken by Barbetti and Colhoun (1988). These specimens were quite strongly magnetised, NRM values ranging from 26.4 to 49.4 μG. All the specimens had an extremely good capability of maintaining a stable remanence. They were cleaned at 35 mT and all had reversed polarity (fig. 11). Twenty eight specimens were taken from silt and clay rythmites by Barbetti and Colhoun (1988), and all of these specimens also had a reversed polarity.

The Gormanston Football Field section records subaqueous deposition in an ice dammed proglacial or supraglacial lake that was subject to ice rafting and occasional subaqueous sediment flows.

Numerous exposures in the Linda Valley show a variety of glacio-lacustrine sediments, most of which exhibit evidence of intense deformation. The field data indicate that the Linda Valley has a history of ice damming by a lobe of the King Glacier flowing from East to West up the valley. During the Thureau advance, ice flowed some 3 km up the valley and formed the large moraine on which the town of Gormanston was built. Although glacial deposits extend up most of the length of the valley and to within 200 m of Karlsons Gap, the saddle that separates Linda Creek from Conglomerate Creek, there is no conclusive evidence to suggest that ice breached the divide and flowed into the Queen Valley.

Regency Formation

Organic deposits of the Regency Formation contain an interglacial flora and overlie Thureau Formation glacial deposits. The relationship between the Thureau and

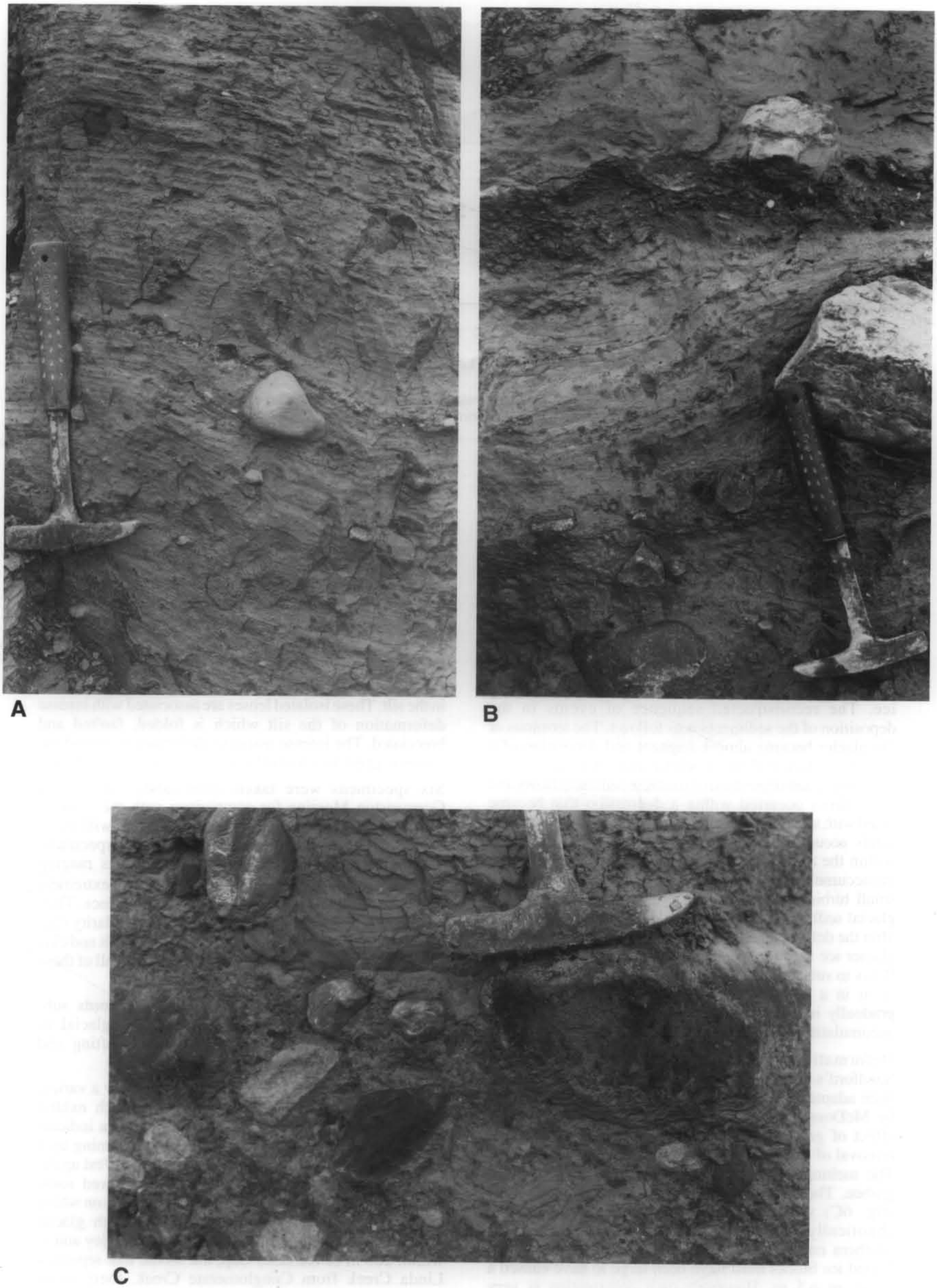


Figure 10. Thureau Formation sediments at Gormanston Football Field. A, Ice-rafted pebbles in gently warped laminated mud. B, Ice-rafted till overlain by a massive debris flow with intraformational blocks of laminated mud, and overlain by faulted laminated mud. C, Detail of deformation at the base of the sediment flow shown in B.

5 cm

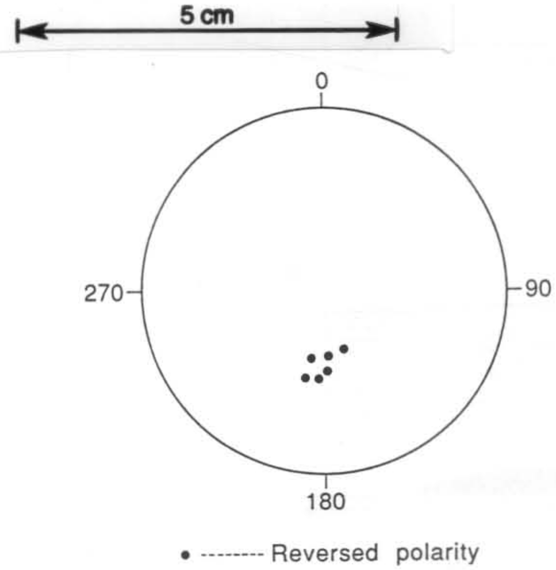


Figure 11. Stereoplots of the detrital remanent magnetisation of the Gormanston sediments.

Regency formations is known from two sections in a quarry in the lower King Valley at CP876312 (fig. 12). In section A-B a lens of organic debris rests conformably on Thureau Formation outwash gravel. It is unconformably overlain by David Formation outwash gravels that appear to be reworked from older till. Pollen analysis of the organic sediments shows an interglacial floral assemblage, which is the Regency Interglacial. Section B-C shows a similar unconformity between massive till of the Thureau Formation and David Formation outwash gravels.

The lowermost sediments of section A-B consist of highly weathered, massive, moderately well-sorted outwash gravel. Pebble fabric of the dip direction of the A/B planes of disc-shaped clasts shows a weak preferred orientation toward 341°, suggesting deposition by a stream flowing approximately from north to south (fig. 12). All Jurassic dolerite clasts in this deposit are completely weathered and form a dull, yellow, clay residue. On the western end of the section a 300 mm-thick dyke of gravel strongly cemented by humic acid precipitate penetrates the host gravel. The dyke can be traced from one wall across the floor into the opposite wall of the excavation. The truncation of this dyke at the top of the gravel demonstrates the presence of a land surface that has now been completely eroded.

Clasts from the highly weathered gravel are mixed with the overlying interglacial organic deposit and suggest that the contact is conformable. This interpretation is supported by the pollen analysis, which suggests succession from a boggy environment with standing water to temperate rainforest. The initial wet environment probably developed on a low-lying, uneven, recently deglaciated surface.

The Regency Formation organic deposit overlies the highly weathered outwash gravel and consists mainly of a combination of *in situ* humified peat matter overlain by drifted wood and leaves. The organic content is very high and ranges from 70 to 95% loss on combustion. The contact with the overlying gravel is sharp, dips west at up to 10° and is eroded. The size of the organic lens appears to have been greatly reduced by erosion.

Pollen analysis and examination of plant macrofossils of the Regency Formation shows an interglacial flora rich in *Lagarostrobos franklinii*, *Nothofagus cunninghamii* and *Phyllocladus aspleniifolius* (fig. 13).

The relative pollen diagram (fig. 13) can be divided into two zones RY2 and RY1 between 350 and 400 mm. The main difference between the zones is the very high percentages of pollen of temperate rainforest trees, particularly *Lagarostrobos franklinii*, above 350 mm, and the greater sclerophyll component indicated by *Eucalyptus* and *Casuarina* below 350 mm. In addition, above 350 mm *Nothofagus* peaks, *Eucryphia – Anodopetalum* and *Anopterus* occur, there are abundant spores of ground ferns and marked peaks of the tree ferns *Cyathea* and *Dicksonia*. These taxa are largely absent below 350 mm where high values of *Pomaderris*, *Athrotaxis*, *Microstrobos* and *Coprosma* occur.

In western Tasmania, substantial quantities of *Pomaderris* usually occur as understorey to wet sclerophyll *Eucalyptus* or in mixed *Eucalyptus* rainforest. *Athrotaxis* is best represented in montane rainforest. *Microstrobos* is a subalpine shrub and *Coprosma* is usually associated with subalpine plants in Tasmanian pollen diagrams.

Although the boundary between RY2 and RY1 is drawn between 350 and 400 mm it is not a sharply defined biostratigraphic boundary; rather it is a transition. The transitional nature is indicated by the reduction of the higher altitude components *Arthrotaxis* and *Microstrobos* at 500 mm while *Eucalyptus* and *Casuarina* do not peak until 400 mm. It is difficult to judge what continuity of pattern in the vegetation changes may have been lost by the absence of pollen from 450 mm.

Two of the most interesting taxa recorded are the extinct species (for Tasmania) *Quintinia psilatispora* and *Gothanipollis perplexus*. Today *Quintinia* is a subtropical wet forest taxon and *Gothanipollis* is a wet forest parasite. *Quintinia* has not previously been recorded from Tasmania in deposits of Pleistocene age. The youngest record is from the fossil soil buried in the Idaho Formation that is believed to be older than the Thureau Formation which is over 730 000 years old. The Idaho record is either of early Pleistocene or late Tertiary age, but the present record is stratigraphically younger than the Thureau Formation. The record suggests that *Quintinia* may have survived in Tasmania until after glaciation had commenced.

The interpretation of the pollen data and the vegetation assemblages that have been inferred indicate that the climate was wet at all times. The large sclerophyll and small subalpine components of RY1 suggest that the vegetation had not yet developed to the optimum condition for the region of temperate lowland rainforest, but that it was either successional or adjusted to cooler montane conditions, or both.

Pollen zone RY2 contrasts markedly with RY1 with the expansion of *Lagarostrobos* to very high values. Tree-ferns and ferns also became important, and the sclerophyll, heath and subalpine taxa decrease markedly. The high values for *Lagarostrobos*, *Nothofagus* and *Phyllocladus* indicate that RY1 represents a lowland temperate rainforest with a marked riparian component. This pollen zone indicates that the organic deposit between the two glacial deposits is not only stratigraphically, but biostratigraphically interglacial, and that it represents part of an interglacial optimum. To what extent the decrease in *Lagarostrobos* in the upper 150 mm reflects the commencement of deterioration of climate and/or soil during the later part of an interglacial cycle, as possibly suggested by the advent of *Eucryphia*, is difficult to say as the sequence has been truncated by meltwater erosion. However, the lower part of RY1 must climatically represent a lowland temperate rainforest environment that was at least as wet and warm as the maximum of the Holocene.

An amino acid date on wood from the organic deposit suggested a minimum age for deposition during oxygen

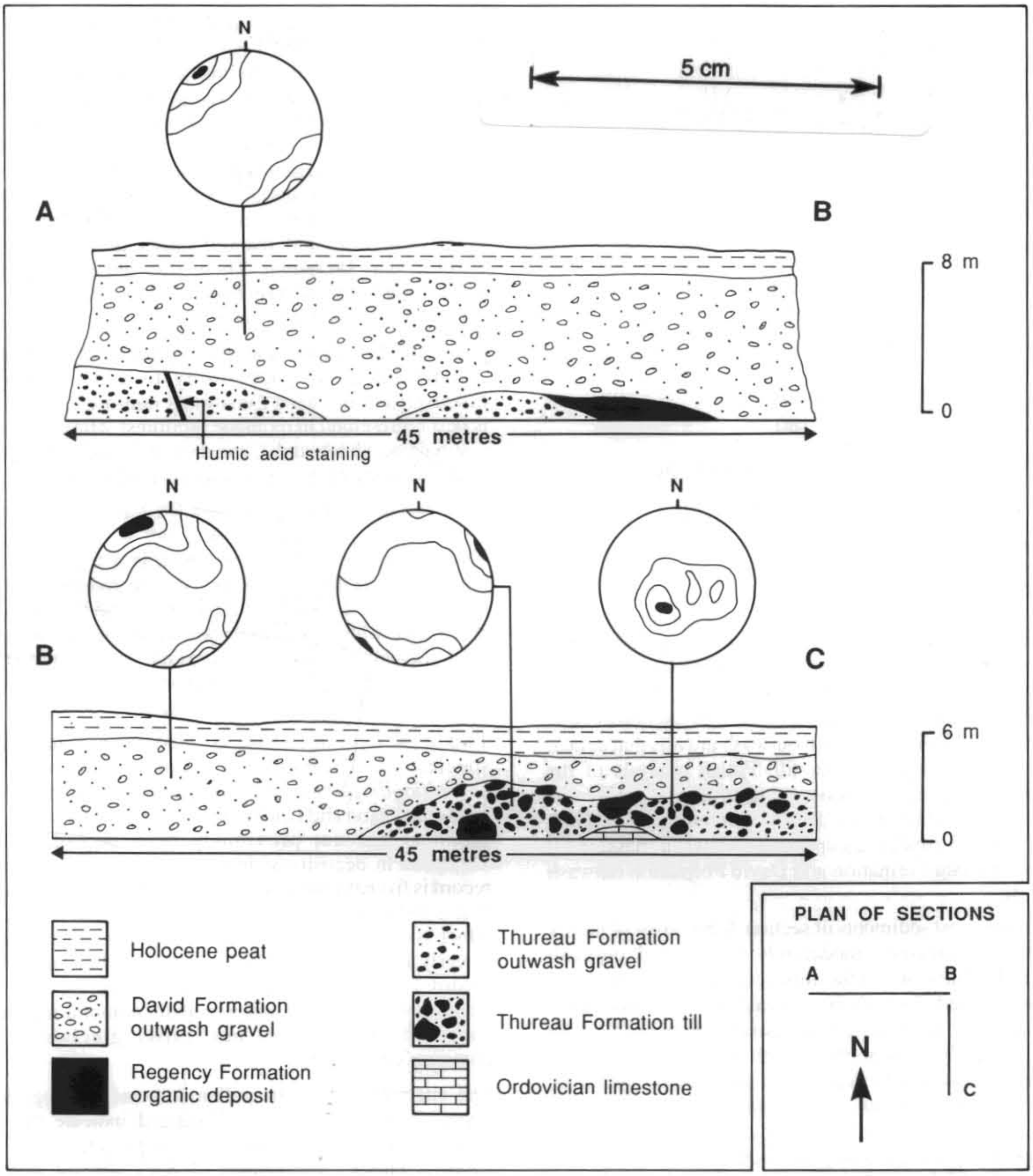


Figure 12. The Regency sections. The pebble fabric of the outwash gravels was measured from the A/B dip direction of disc-shaped pebbles and for the diamictons the A axis dip direction was measured. The equal area nets are contoured at a 2 standard deviations interval using the method of Kamb (1959).

isotope stage 8 (B. J. Pillans, pers. comm., 1987). This date conflicts with the local relative stratigraphy which suggests the Regency Formation is considerably older. This apparent conflict has not been resolved and is discussed further in the section on dating.

Section B–C is oriented north to south and intersects the eastern end of the Regency section described above at right angles (fig. 12). The basal sediments of this section include a massive, highly weathered till that rests on a light grey massive clay. This clay appears to be part of the weathered Ordovician limestone that underlies most Quaternary sediments in this area. Erratic boulders of Jurassic dolerite, Ordovician conglomerate and Permian sediments up to 1.5 m in diameter are common in the till. Weathering rinds on Jurassic dolerite clasts have a mean thickness of 75.5 mm and a standard deviation of 14.5 mm. The pebble fabric of the diamicton is weak and is unrelated to the reconstructed ice flow direction (fig. 12). The right hand fabric diagram shows that the clasts have a very steep dip, which is not characteristic of any known origin for glacial deposits. It may be due to localised deformation.

The contact with the overlying gravels is uneven, sharp and eroded. Where large weathered dolerite boulders protrude through the till their weathering crust has been removed by scour associated with the currents that eroded the till and redeposited it as gravel.

The till is overlain by coarse, poorly-sorted gravel with particle sizes of up to 800 mm in diameter. It contains occasional lenses of sorted sand and resembles the upper gravel in section A-B. The gravel grades upward into poorly-sorted sand which is overlain by up to 600 mm of fibrous peat. The surface of the deposit is a relatively flat terrace marked by several sinuous channels that have a braided pattern. The surface of the terrace is marked by several clusters of large erratic boulders up to 2.5 m in diameter which appear to be lags from the erosion of Thureau Formation tills.

Comparison of the two Regency sections shows that the Thureau Formation tills in section B-C are 1 to 2.5 m above the Thureau Formation outwash gravels in section A-B. This indicates that the till protrudes through outwash gravel. The pebble fabric suggests that the till has been

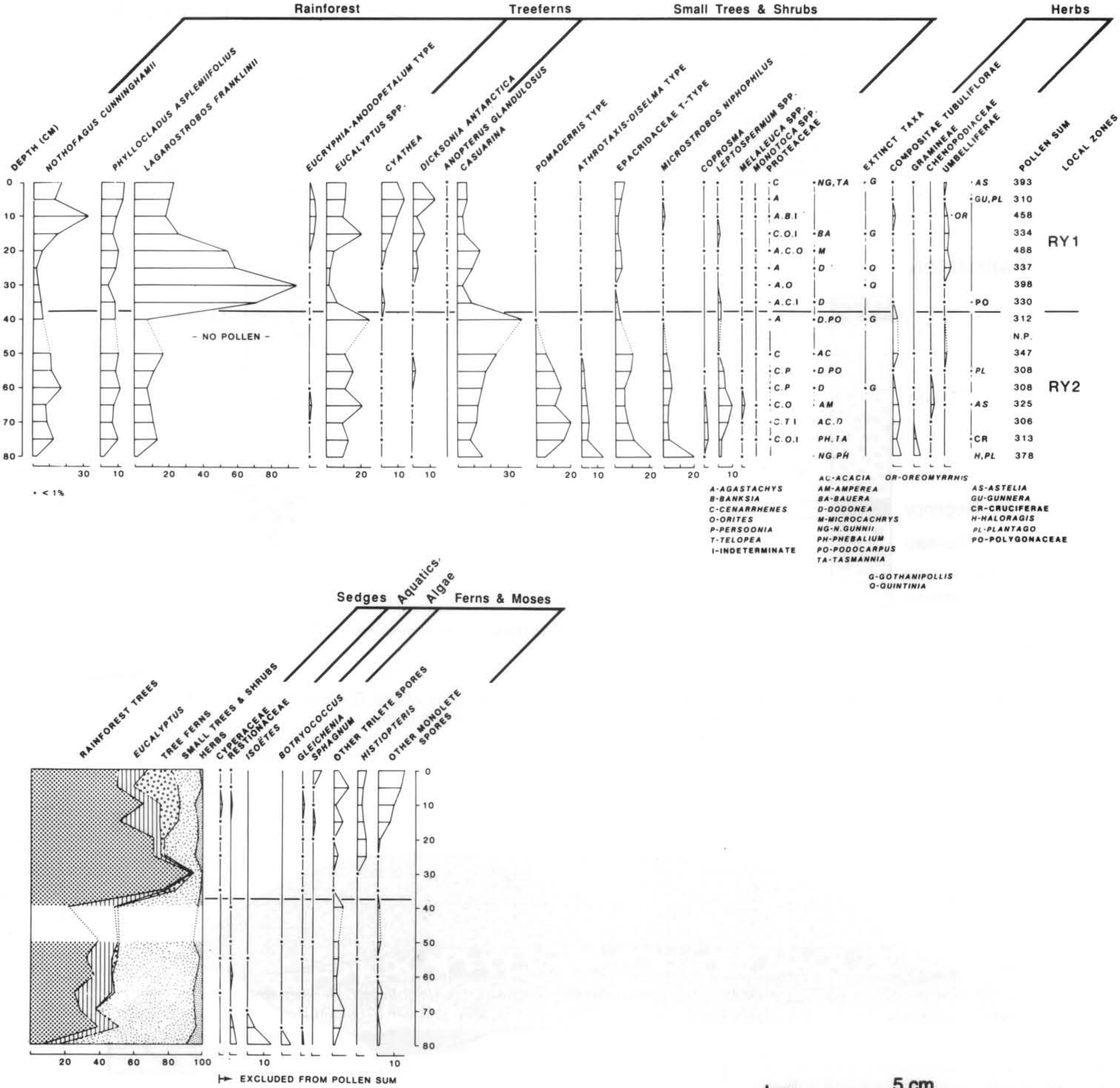


Figure 13. Pollen diagram of the organic sediments of the Regency Formation.

significantly modified by secondary processes after its release from the ice as the orientation of clasts bears no relationship to the direction of ice movement (fig. 12). Comparison of the fabrics with those for secondary processes observed by Lawson (1979) suggests the first fabric is more typical for sediment flows and the second may be some kind of dump deposit that has accumulated as ice slope colluvium by falling or sliding off an ice surface. The former alternative of deposition in a supraglacial position is considered the more likely of the two depositional environments.

The organic sediments that lie above the Thureau Formation sediments in the section record the only pollen evidence of interglacial warming between the Thureau Formation and younger sediments. The geometry of the organic lens suggests that it was formerly of greater extent and has been eroded by sediment fluxes associated with the succeeding glacial advances.

The interpretation of the overlying sediments as David Formation outwash gravel is based on the recognition of the aggradation surface being inset into adjacent, topographically higher Cableway Formation outwash gravels. An interpretation of the stratigraphic relationships across the valley in this area shows the relationship between the David, Cableway, and Thureau Formations and the Ordovician limestone (fig. 14).

Huxley Formation

The Huxley Formation is known from remnants of outwash gravels that rest topographically above the aggradation level of the Pyramid Formation. Sediments

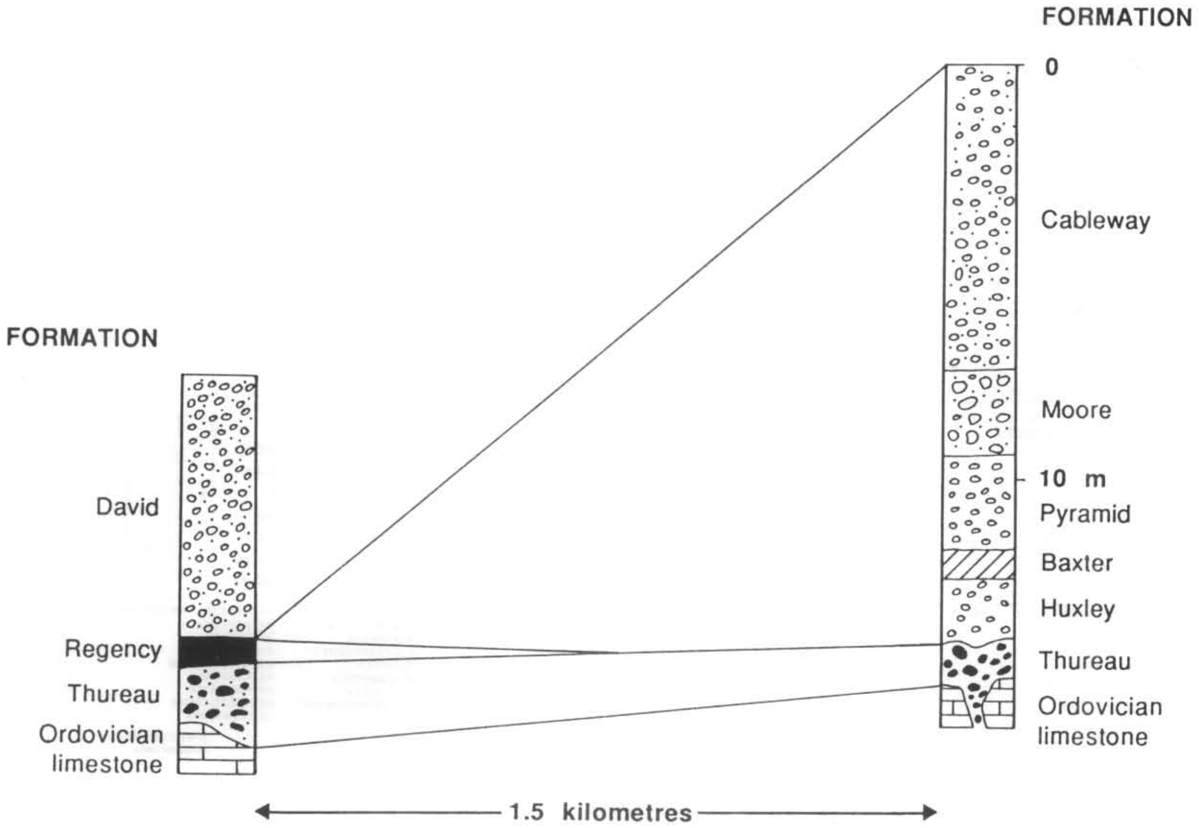


Figure 14. Reconstructed relationship of the contacts between the Cableway, David, Moore, Regency and Thureau formations in the lower King Valley as observed at the Regency site (left) and the Baxter Rivulet site (right).

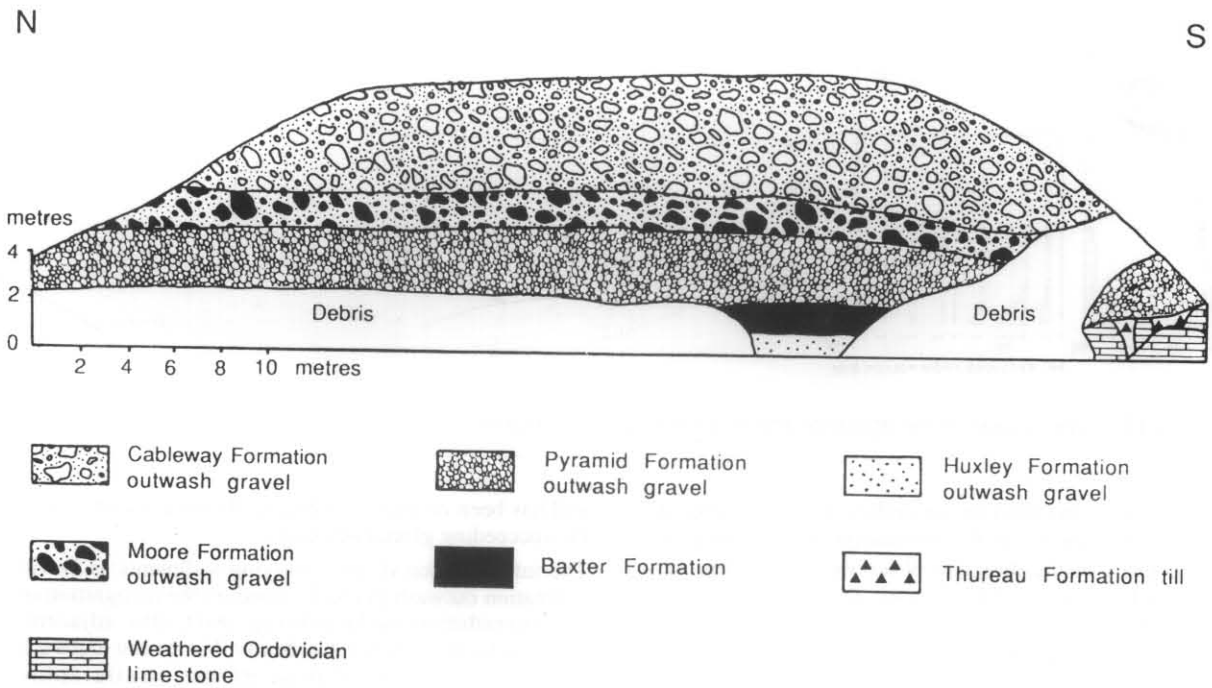


Figure 15. The Baxter Rivulet section.

5 cm

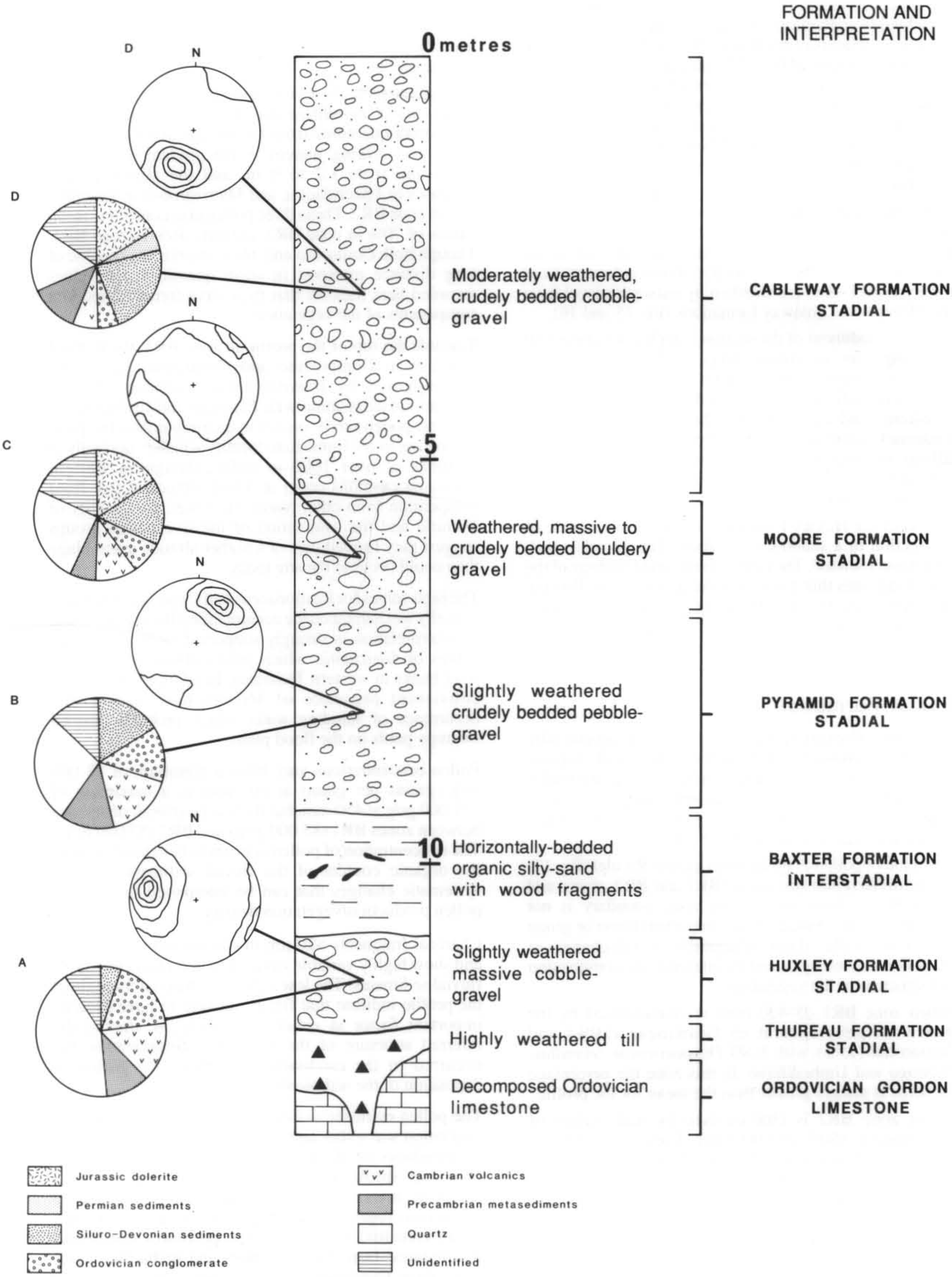
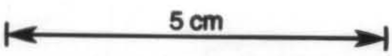


Figure 16. Generalised lithological log with fabric and lithological data of the main stratigraphic units exposed at Baxter Rivulet. Equal area projections, contoured at an interval of 2 standard deviations using the method of Kamb (1959).



of this formation are not very extensive in the King Valley and are difficult to distinguish from sediments of the Pyramid Formation. However, at Baxter Rivulet they are separated from Pyramid Formation sediments by 1.2 m of organic sediments of the Baxter Interstadial. Because the sediments are derived from Mt Jukes their extent is limited to the middle King Valley. The extent of the ice advance associated with the sediments is not known, but it is thought that it may have breached the Baxter Rivulet catchment and entered the Governor River. Lack of exposure in the Governor River make this impossible to confirm.

The type section exposed on the right bank of Baxter Rivulet at CP875300 shows Huxley Formation outwash gravel resting beneath organic silty sands that record an interstadial flora. The same section shows sediments from the Pyramid Formation overlain by outwash gravel from the Moore and Cableway formations (fig. 15 and 16).

The basal sediment of the section is highly weathered till that crops out as narrow dykes in the surface of the weathered Ordovician limestone (fig. 15). Boulders of Jurassic dolerite up to 150 mm in diameter are completely weathered and the till is thought to correlate with the Thureau Formation which has been observed to occur as fillings of solution pipes elsewhere in the valley. The contact between the till and Huxley Formation gravels was not observed at this section.

Gravel of the Huxley Formation is seen in the middle of the section in a shallow excavation. The gravel is well sorted and rounded. The pebble fabric and lithology of the gravel indicates that it was deposited by a stream flowing from Mt Jukes (fig. 16). The Huxley Formation outwash gravel is overlain by 1.2 m of organic silty sand that records an interstadial palynoflora of the Baxter Formation.

Baxter Formation

The Baxter Formation consists of 1.2 m of organic silty sand. The deposit is horizontally bedded and contains drifted wood fragments up to 60 mm long (fig. 16) Pollen analysis of the silty sand shows an herbaceous assemblage with an alpine component. The type and only section occurs at Baxter Rivulet at CP875300.

The relative pollen diagram from Baxter Rivulet (fig. 17) can be divided into two zones, BR1 and BR2, above and below 0.4 m respectively. The zone boundary is not well-defined by changes in any important taxon or group of taxa, but is placed having regard to overall changes in the pollen assemblages, and by inference the composition and structure of the vegetation.

Pollen zone BR1 (0–400 mm) is characterised by the maximum development of Gramineae (19%) and Compositae (15%), with 3–8% *Donatia novae-zelandiae*, *Plantago* and Umbelliferae. In this zone the percentage for herbs is always greater than the mean for the profile.

Pollen zone BR2 is characterised by high values of Epacridaceae (50% at 500 mm), *Casuarina* (29% at 1.2 m) and *Microstrobos* (20% at 700 mm). In contrast to zone BR1, tree and alpine and subalpine shrub pollen values are greater than the mean values of 20% and 11% for the profile respectively, except below one metre where alpine and subalpine pollen decrease below 9%.

The two major temperate rainforest taxa *Nothofagus cunninghamii* and *Phyllocladus aspleniifolius* never exceed 8% collectively, and generally do not exceed 4%. If temperate rainforest was present these values would almost certainly exceed 25%. The values for *Lagarostrobos franklinii* are also generally low and only exceed 10% at the base of the profile. These low values suggest the presence of occasional trees or small stands of trees close to the river bank. The very low values of *Dicksonia antarctica*, which do not exceed 0.3% on any horizon, also suggest that all taxa usually found in or adjacent to rainforest were near their upper limits of distribution or were absent from the area. Although *Eucalyptus* pollen is more likely to be derived locally than that of rainforest taxa, the low values not exceeding 5% suggest there were few eucalypts.

Most of the pollen is derived from Epacridaceae (T-type), *Casuarina* cf *monilifera* and *Microstrobos niphophilus* with small quantities from Gramineae and Compositae. The most striking pattern is the gradual decrease in *Casuarina* from 1.2 to 0 m, and the slight relative increases of Epacridaceae and *Microstrobos* in the upper part of zone BR2. These three pollen taxa contribute mean values of 70% in zone BR2 and only 46% in zone BR1. Though both *Casuarina* and *Microstrobos* are capable of long distance transport in small quantities, the values recorded here indicate that these taxa formed important components of the vegetation.

Towards the top of the profile in zone BR1, the marked increase in Compositae and Gramineae pollen are accompanied by *Donatia novae-zelandiae* (8% at 300 mm) in association with *Plantago* and Umbelliferae. *Donatia* pollen rarely occurs in quantity unless the plant is growing in the immediate locality which is normally at altitudes of over 1000 m today, though in western Tasmania cushions occur at lower altitude if free from competition from other plants. However, the quantity of *Donatia* and the association of the five pollen groups suggest they formed part of a higher altitude assemblage than could occur at the site today.

The high values for Restionaceae and Cyperaceae indicate a locally wet habitat on the accumulating flood plain. Such an interpretation is strongly supported by the very high values of *Gleichenia* which grows abundantly on low river banks in western Tasmania. In addition, the almost consistent presence of *Myriophyllum* shows the occurrence of standing water which probably formed swampy pools on the flood plain.

Pollen concentrations vary from a minimum of 33 000 gr/g (grains per gram) at 100 mm to a maximum of 307 000 gr/g at 450 mm, but there is no overall difference between zones BR1 (85 000 gr/g) and BR2 (87 000 gr/g). The concentration of pollen appears to be related more to the organic content of the fluvial sediments than to systematic changes that can be interpreted in terms of pollen production/vegetation density.

Charcoal fragments >20 μm in size occur in all samples and show highly variable values which is characteristic of fluvial sediments. The low values of charcoal throughout the profile indicate that fire is unlikely to have been an important factor in either establishing or altering the inferred structure of the vegetation communities, but occurred in the catchment throughout the period of deposition of the sediments.

The pollen evidence indicates that during zone BR2 the vegetation was either an open *Casuarina* woodland with an abundance of *Microstrobos* and Epacridaceae shrubs or a wet shrub-heathland in which *Casuarina* was abundant. It was neither a forested nor a predominantly herbaceous vegetation. The vegetation became more open during zone BR1 as the average proportion of woody taxa was reduced from 86.4% to 60% and herbaceous taxa, particularly Gramineae and Compositae, increased from 13.6% to 40%. The change suggests that either the understory of the woodland became grassy or that a mosaic of scrubland-herbland-heathland vegetation was developed.

The preferred interpretation is that the vegetation was a wet *Casuarina* heath which either became a herbland-heath mosaic or a mix of herb and heath species. This inferred vegetation differs from other non-forest

BAXTER RIVULET

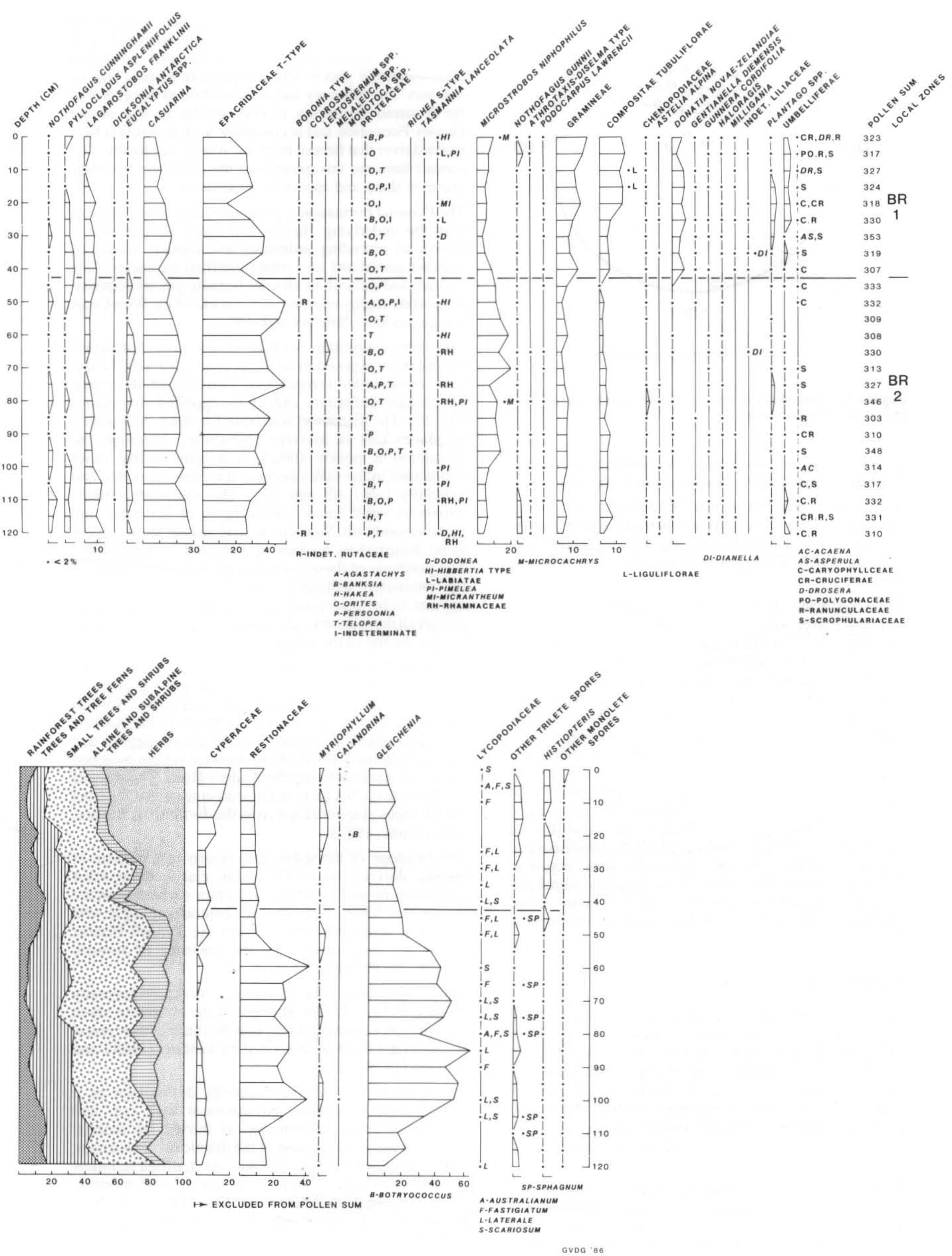


Figure 17. Baxter Rivulet pollen diagram.

5 cm

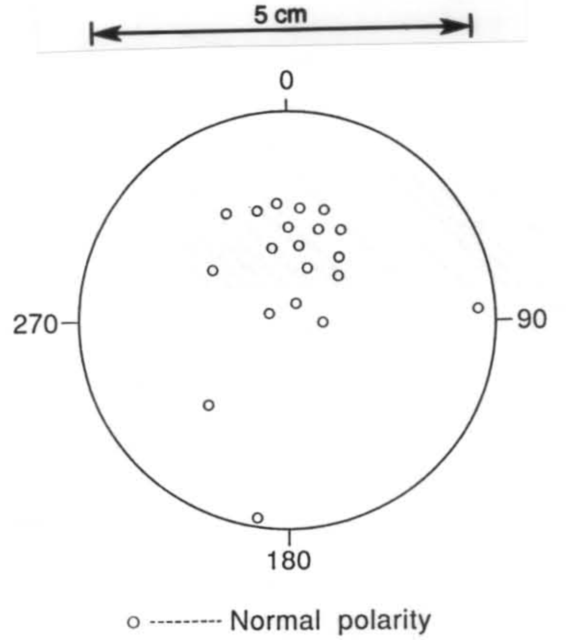


Figure 18. Stereoplot of the detrital remanent magnetisation of the Baxter Rivulet silts.

pollen vegetation assemblages recorded from lowland western Tasmania in the abundance of *Casuarina*.

The flora is non glacial and lies between two glacial outwash gravels. It records an interstadial event *sensu stricto*, that is a non glacial period in a glacial event (Bowen, 1978). The interstadial is known only from this site.

Dates on wood in the overlying Cableway Formation outwash gravel are considered to be infinite because they are close to ^{14}C background (table 4). No attempt was therefore made to carbon date wood found in the Baxter Formation. The Baxter Formation is certainly beyond ^{14}C age and is probably of middle Pleistocene age. Wood from the Baxter Formation gave an amino acid age that suggests a minimum age equivalent to oxygen isotope stage 10 (B. J. Pillans, pers. comm., 1987). Twenty specimens were taken from the organic silty sand for palaeomagnetic analysis. All the specimens were weakly magnetised (NRM values ranging from 0.079 to 0.547 μG). Three quarters of the specimens had a fair capability of maintaining a stable remanence, whilst the remainder had a poor capability. All the specimens were step-wise demagnetised because of their weak magnetic signal and the very low median destructive field of many of the specimens. All the specimens had a normal polarity (fig. 18).

Pyramid Formation

The Pyramid Formation was deposited by an ice advance of the Jukes Glacier, the type section is the same as that of the Baxter Formation. At this section the Pyramid Formation is represented by 4 m of coarse outwash gravel, and is overlain by 1.6 m of Moore Formation outwash gravel (fig. 16). Another important section of Pyramid Formation gravels occurs at CP870274 adjacent to Fish Creek, and outwash gravels extend discontinuously to the Andrew Divide.

The Pyramid Formation, like the Huxley Formation, is a suite of glacigenic deposits derived from Mt Jukes. It is limited in extent to the area south of the Governor River in the King Valley. Except at Baxter Rivulet, the Pyramid Formation unconformably overlies, and is indistinguishable from the Huxley Formation. The ice-contact sediments of this formation form relatively rare isolated low hills within low-lying outwash surfaces. The type section of the Pyramid Formation occurs at the Baxter Rivulet section at CP875300 that was described previously.

The Pyramid Formation outwash gravel overlies the Baxter Formation organic sediments and consists of up to 1.7 m of massive, well-sorted and rounded clast-supported gravel with a maximum particle size of 600 mm. The lithology and pebble fabric of the gravel of the Pyramid Formation is very similar to that of the Huxley Formation and is consistent with deposition by a braided river that flowed from the West Coast Range. The contact between this gravel and the overlying bouldery gravel is sharp and appears to be scoured.

The Pyramid Formation is geographically more extensive than the underlying Huxley and Baxter formations. It covers all preceding sediments and consists mainly of outwash gravel and ice-contact stratified sediments. Tills are rare and difficult to identify because glacial comminution of the highly resistant source rocks of the West Coast Range produces little silt and clay.

The other important section of Pyramid Formation sediments is exposed in an excavation at CP870274 in a low, hummocky terrace that consists of deformed ice-contact sediments and clast-supported diamictons (fig. 19). The diamicton is at least 3 m thick, is stratified in places and is massive elsewhere. Although the diamicton is poorly sorted, it is quite unlike other diamictons in the valley because it is clast-supported and generally lacks silt and clay. The sediment contains numerous small lenses and stringers of laminated and massive sandy silt. Some of the sand and silt lenses appear to be disturbed by a series of high-angle reverse faults. The positions of these faults are inferred from offsets in the laminated sediments (fig. 19). The deformation appears to be syndepositional because the faults do not disturb all the sediments and appear to increase in intensity toward the top of the section.

Pyramid Formation outwash gravel also extend southward to the Andrew Divide where up to 5 m of gravel rests on weathered Ordovician limestone. Similar outwash gravel occurs beyond the divide in the Andrew River catchment. The location of the gravel suggests that outwash streams from the King drainage system breached the divide and flowed into the Andrew River. The absence of till and large boulders in the area suggest that ice from the Mt Jukes glacier did not cross the divide (F. J. Baynes, pers. comm., 1987).

The lithology of the sediments consists of a mixture of locally derived Siluro-Devonian clasts, Ordovician conglomerate and Cambrian volcanics derived from the West Coast Range. The lithology contrasts strongly with overlying sediments derived from the Tyndall Plateau which are distinguished by the presence of Jurassic dolerite and Permian sediments.

The pebble fabric of the diamicton is weak, and generally unimodal (fig. 20). The strength of the clustering about the principle eigenvector (S_1) is in the range of that typical for sediment flows as described by Lindsay (1968), and Lawson (1979).

The Fish Creek section records deposition in a supraglacial ice-contact environment with syn- and post-depositional deformation as buried ice melted. Although the precise origin of the diamicton is uncertain, the intraformational lenses of silt and stratified sediment resemble subfacies 1, facies A of supraglacial morainic till as described by Boulton and Eyles (1979). Such an association of deformed sediments is typical of deposition at the frontal margin of valley glaciers during retreat.

Silt and mud lenses, and stringers are common in supraglacial tills. They form as mud is eluviated from debris overlying ice and deposited at the ice-sediment interface. The position and geometry of silt stringers in

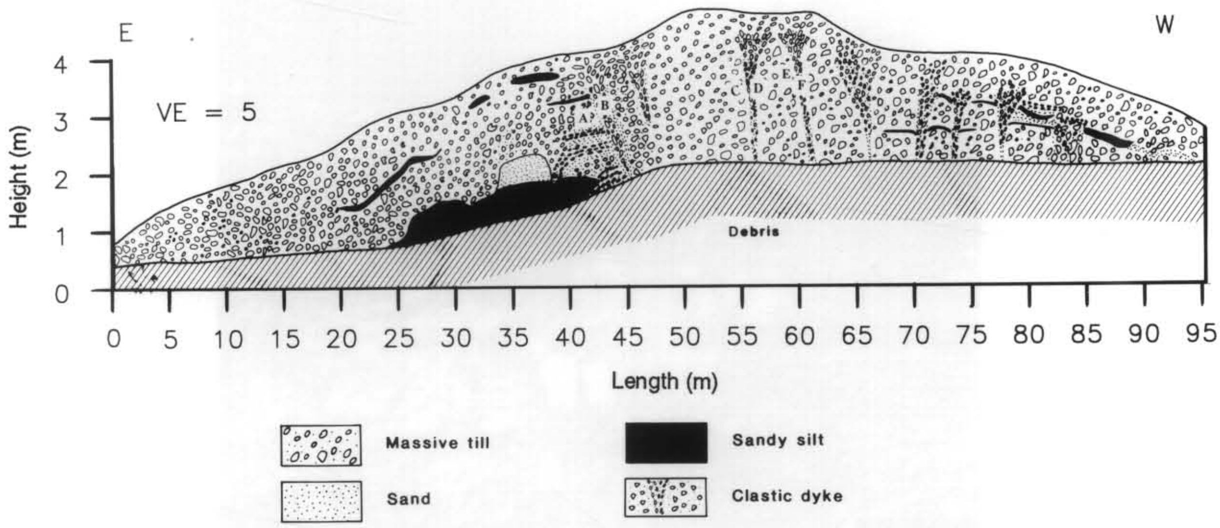


Figure 19. Supraglacial sediments of the Pyramid Formation exposed near Fish Creek. Letters correspond to fabric analyses shown in Figure 20.

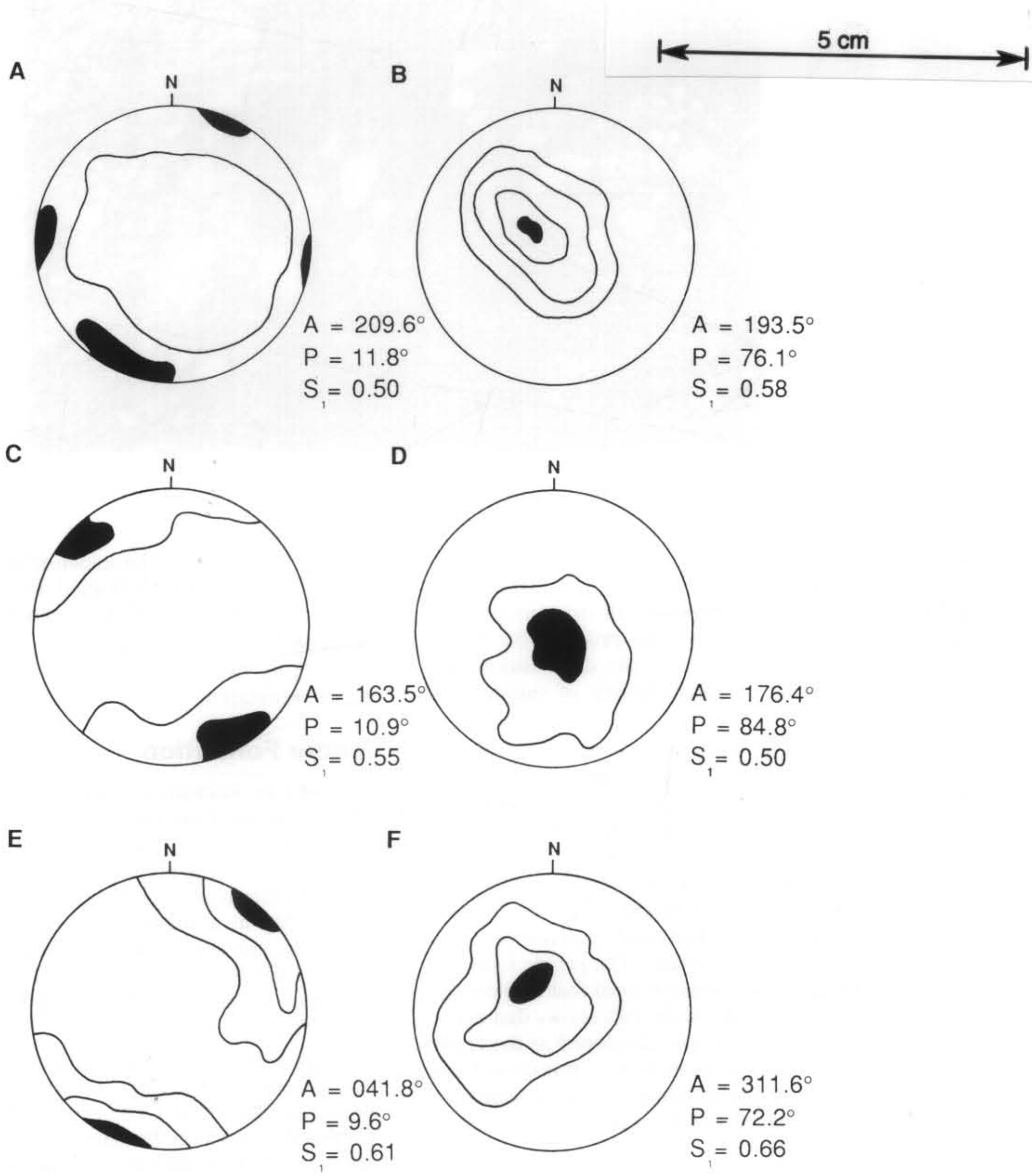


Figure 20. Pebble fabric of the Pyramid Formation supraglacial sediments. The equal area nets are contoured at 2 standard deviation intervals using the method of Kamb (1959). The zone of maximum clustering is black. See Figure 19 for sample locations.

Figure 21. Large straight sided clastic dyke in Pyramid Formation sediments. 5 cm

the section suggest there are two or three ice-sediment interfaces recorded in this section (fig. 19). During the final melt of the underlying ice the silt stringers become increasingly deformed during the process that Eyles (1979) called backwasting. Both the deformation and multiple silt stringers are characteristic of supraglacial sediments.

The section has several wedge-shaped clastic dykes and collapse structures that penetrate the diamicton by up to 3 m. (fig. 19). These dykes have been described by Fitzsimons and Colhoun (1989), where this group is referred to as swarm 2. The dykes are narrow, straight-sided and wedge-shaped structures, and many have walls made of vertically-dipping coarse sand (fig. 21). The fills appear to consist of reworked host material that has slumped or fallen into open cracks. The pebble fabric of the host sediment is very different from that of the dyke fills (fig. 19, 20). The fabric of the fills shows that clasts have a vertical dip that probably developed as the crack opened and filled. There appears to be a transition from wide and shallow collapse structures on the western side of the section to the narrow and deeper dykes in the central part of the section (fig. 19).

The origin of the clastic dykes is not certain but they seem to be very similar to boulder-filled tension cracks that have been observed to form during the process of backwasting on Icelandic glaciers (Eyles, 1979, p. 1345). This process appears to be very similar to deformation in glaciofluvial gravels described by McDonald and Shilts (1975) using Sandford's (1959) model deformation experiments.

At Baxter Rivulet Pyramid Formation outwash gravel is overlain by 1.6 m of bouldery gravel that is part of the Moore Formation.

Moore Formation

The Moore Formation, which was deposited by an advance of the King Glacier, is known to occur only in two exposures near Baxter Rivulet. It has no surface expression and it is not represented on the map showing the distribution of stratigraphic units. The deposits are found only in a small area between the limits of the Cableway advance and the southern limit of erratics (fig. 1). The formation has no surface form because the sediments are eroded and buried by Cableway Formation outwash gravel.

The type section of the Moore Formation is an excavation 200 m south of the Kelly Basin Road bridge over the Governor River at CP881303. The deposits of the Cableway and Moore formations are easily distinguished on the basis of weathering of Jurassic dolerite clasts. At the type section the Moore Formation consists of 4.3 m of massive, coarse gravel with particle sizes up to 300 mm in diameter that is overlain by finer gravel interbedded with gravelly sand (fig. 22). All the gravel is moderately well sorted, has numerous voids, is diffusely iron-stained to a bright orange colour and is locally iron-cemented. The lithology of the gravel consists of a mixture of clasts that have a West Coast Range, Eldon Range or local provenance (fig. 22).

The differences in weathering rind thicknesses on Jurassic dolerite clasts suggests that a substantial period of time separates the deposition of the Moore and overlying Cableway Formation. The Moore Formation gravels have weathering rinds with mean values of 14.3 and 17.3 mm compared to 5.3 and 5.1 mm in the Cableway Formation (fig. 23). The contact dips west at up to 10° and is marked by the protrusion of partly eroded larger clasts through the contact and by dish-shaped load structures in a thin bed of fine silty sand. The deformation structures in the silt suggest that it was deposited during the onset of the Cableway advance and was subsequently loaded by the gravel. Both gravels are penetrated by steeply-dipping zones of humic staining from the overlying Holocene peat. The process of leaching of humic acids and deposition in vertical zones in sediments is common in the King Valley.

At Baxter Rivulet the Moore Formation consists of coarse, poorly-sorted gravel with boulders up to 700 mm in diameter. The lithology suggests the gravel is a deposit from the King Glacier (fig. 16). The pebble fabric suggests the gravel has been derived and redeposited from a till (fig. 16). The contact between these gravels and the overlying outwash sediments of the Cableway Formation is sharp, and from the protrusion of larger boulders through the contact it appears to have been scoured. The contact between the Pyramid and Moore Formation sediments can be traced southward where it manifests itself as a lithological boundary on the surface of the terrace. This boundary, which can be traced across the floor of the King Valley, is the southern limit of erratic dolerite and Permian clasts from the King Glacier.

The Moore Formation gravels form an upward-fining sequence of braided river outwash gravels deposited in an ice proximal position. Lack of structure in the lowermost gravels suggests that they were deposited by a sheet flood. Because the texture of the gravel is coarser than the Cableway Formation deposits they are thought to have been deposited close to the ice source. Although there is no surface form to this deposit the texture of the gravels at this section and at Baxter Rivulet suggest that the ice-terminal position was in a similar location to that of the Cableway Formation (fig. 2).

Cableway Formation

The Cableway Formation was deposited by an extensive advance of the King Glacier. During this advance ice flowed 17 km down the King Valley and terminated near the King River bridge where a small moraine and outwash terrace are preserved at CP886314. The formation consists mainly of terraces of outwash gravel with occasional lenses of buried till and a few low hills of ice-contact stratified deposits.

Outwash sediments of the Cableway Formation form extensive terraces that extend from the Thureau Hills to the Crotty Plain, where the glacier terminated (fig. 2). Because the ice advance terminated in a relatively narrow part of the valley, many of the ice-contact sediments have been eroded by outwash streams of the same and subsequent ice advances.

The type section of the Cableway Formation is a 28 m roadside exposure near the King River bridge at CP886314. The exposure is in a small terrace remnant high on the valley side that rests on Siluro-Devonian quartzite. Sediments in the exposure record an upward-fining sequence of proximal outwash gravels with large cut and fill structures that are overlain by locally derived slope deposits (fig. 24).

The lithology of the outwash gravels is a mixture of locally derived and erratic rocks although the gravels have a much lower erratic content than other glacigenic sediments in this part of the valley. They have a variety of structures including horizontal bedding, cross-bedding, large cut and fill structures and old channel fills (fig. 24). All the gravels are moderately well sorted, and are interbedded with sands and sandy gravels. They rest on an uneven Devonian quartzite surface that is highly abraded and smoothed in some areas, and highly angular in others. Large angular blocks of up to one metre diameter have

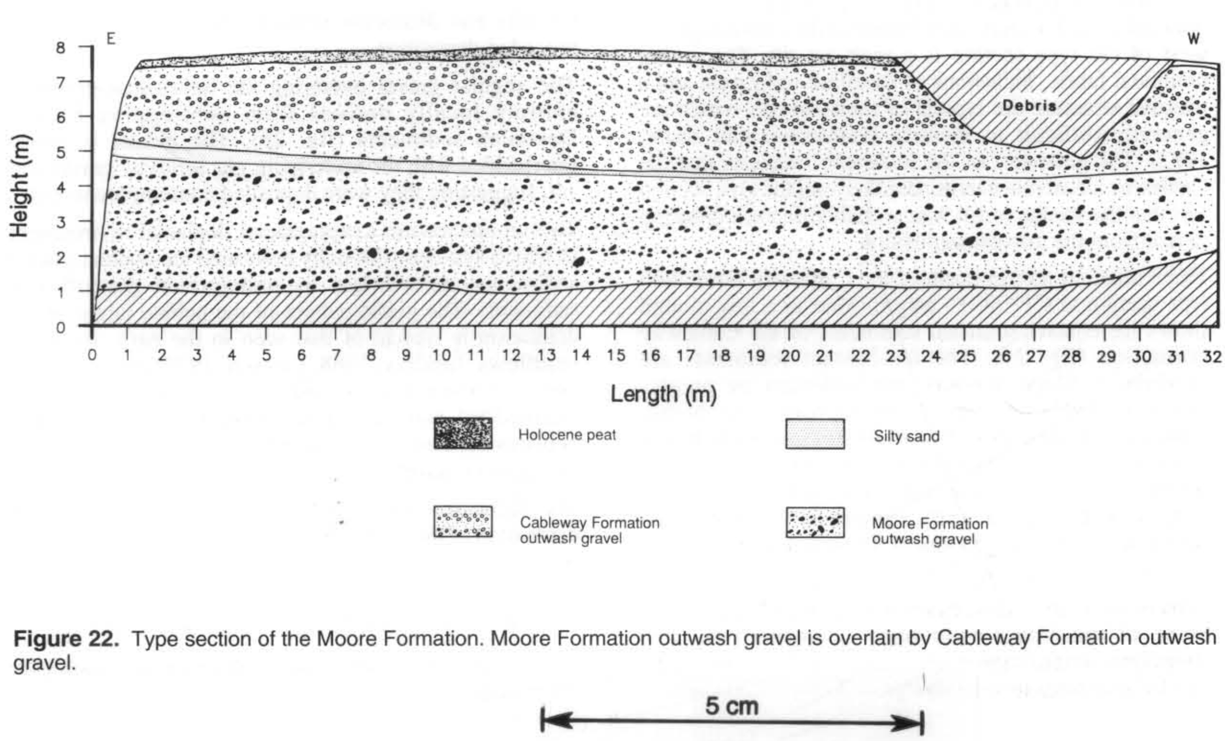


Figure 22. Type section of the Moore Formation. Moore Formation outwash gravel is overlain by Cableway Formation outwash gravel.

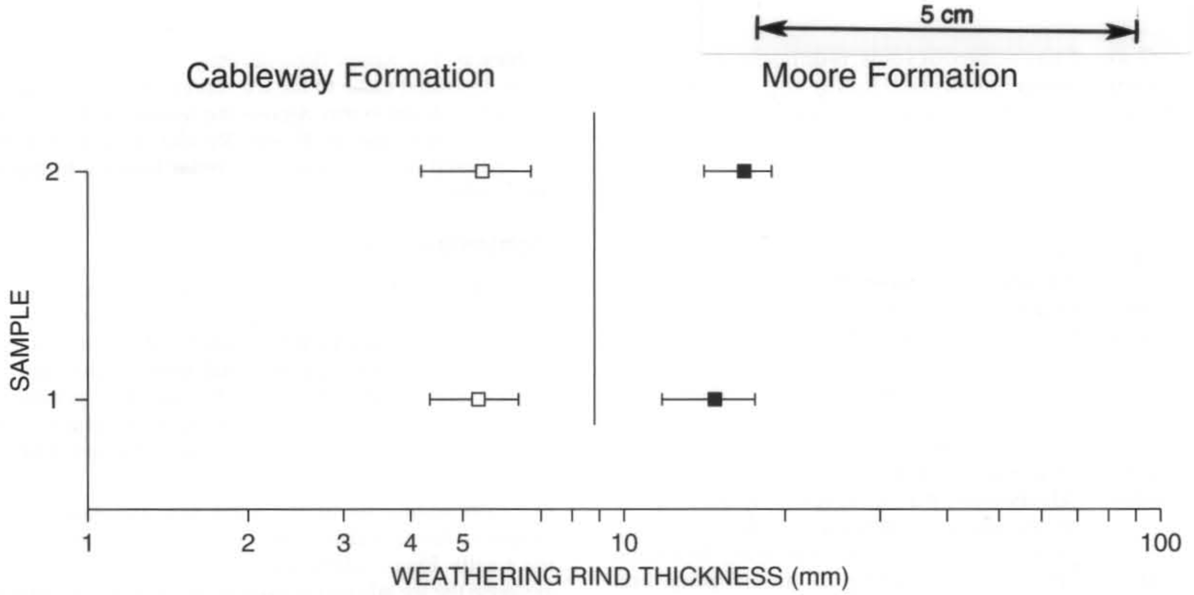


Figure 23. Differences in weathering between the Cableway and Moore Formations. Each point represents the mean of 50 measurements and the bar represents ± 1 standard deviation.

been incorporated into the gravel in places (fig. 24). Up to 11 m of locally derived slope deposits overlie the outwash gravel and consist of angular clasts of Devonian quartzite with a sandy matrix. (fig. 24). Although the quartz sand is well sorted and has the appearance of aeolian sands it may be the product of physical weathering of the Siluro-Devonian Quartzite.

Sixty metres west of the type section a series of sections show a northward-thickening wedge of diamicton that lies between the outwash gravel and bedrock (fig. 24). The diamicton is mainly massive but includes some semi-stratified zones that do not form discrete lenses. Both the weathering and lithology of the diamicton contrast with that of the outwash gravel. The weathering rinds on Jurassic dolerite in the till have a mean thickness of 7.6 mm and a standard deviation of 1.8 mm. The mean weathering rind thickness of clasts in the gravel is 4.7 mm and the standard deviation is 1.1 mm. The diamicton is considerably richer in erratic Jurassic dolerite and Permian sediments than the outwash gravel.

The thick gravel deposit is ice-proximal outwash that was deposited at the maximum of the Cableway ice advance. The large cut and fill structures are from unstable stream beds that are typical of steep ice-proximal reaches of outwash fans. The diamicton buried in the outwash gravel west of the type section is a melt-out till. The strong unimodal fabric, inclusions of stratified sediments and a massive structure are typical of melt-out tills which are generally difficult to identify (Haldorsen and Shaw, 1982). The geometry of the sediments suggests that the Cableway ice advance terminated beyond the type section and that the wedge of till was probably deposited during retreat from the maximum position.

Three hundred metres north of the Cableway Formation type section at CP878313 an exposure in a low mound shows ice-contact stratified sediments of the Cableway Formation (fig. 25). The ice-contact sediments are overlain by slope deposits and underlain by poorly-exposed, highly weathered till deposits. The highly weathered till contains Jurassic dolerite clasts up to 40 mm in diameter that are completely weathered to clay. The pebble fabric of the diamicton shows a weak unimodal fabric dipping toward 122°. The gravel is probably an eroded remnant of the Moore or Thureau formations.

The till is unconformably overlain by a series of deformed gravels, silts and diamictons with complex geometry. Some of the silts are intensely deformed (fig. 25). Numerous small, high-angle reverse faults occur in the bedded sediments in the western side of the section. Sandy silts rest on a coarse diamicton on the eastern side of the section and on bedded gravels on the western side. The massive diamicton on the eastern side of the section is up to 1.5 m thick, contains lenses of poorly-sorted, silty sand and has clasts up to 200 mm in diameter. It is clast-supported in places and matrix-supported in others. The pebble fabric of the clast-supported part of the diamicton has a weak girdle pattern (fig. 26B). The stratified gravel, which lies above and against the diamicton on the western side of the section, is poorly sorted, crudely bedded, imbricated, and contains clasts up to 100 mm in diameter. The pebble fabric of the stratified gravel suggests it was deposited by a current that flowed from the north-east to south-west (fig. 26C).

Overlying the deformed diamicton, are well sorted, relatively undeformed, horizontally bedded gravel and silt deposits. The pebble fabric of the fine gravel shows a strong unimodal concentration very similar to that of the underlying imbricated gravel (fig. 26A). Gentle warping of the bedding on the western side of the section suggests some degree of syn-depositional deformation. However, as the remainder of the gravel is undeformed it was probably deposited after the main period of deformation of the underlying sediments. The sharp contact with both the silts and diamicton suggests that a period of scour preceded deposition.

The angular gravel overlying the imbricated gravel is entirely locally derived from outcropping Siluro-Devonian sediments (fig. 25). The gravel is partly interbedded with the underlying glaciofluvial gravel and silt, suggesting they were deposited contemporaneously.

This section records a supraglacial, deglacial environment in which syn-depositionally deformed ice-contact glacial sediments are overlain by undeformed glaciofluvial sediments. The girdle pattern of the pebble fabric of the diamicton is typical of that seen in the early stages of mudflows (Lindsay, 1968; Lawson, 1979) and may have been deposited as a slumped or flowed till. The interbedded and deformed nature of the relationship between the diamicton and outwash gravels is typical of supraglacial sediments.

The reconstructed sequence of events leading to the deposition of the sediment association seen at this section is:

(1) melting of the glacier which terminated about 50 m from the section during the Cableway ice advance;

(2) deposition of sand lenses from englacial tunnels onto the melting ice;

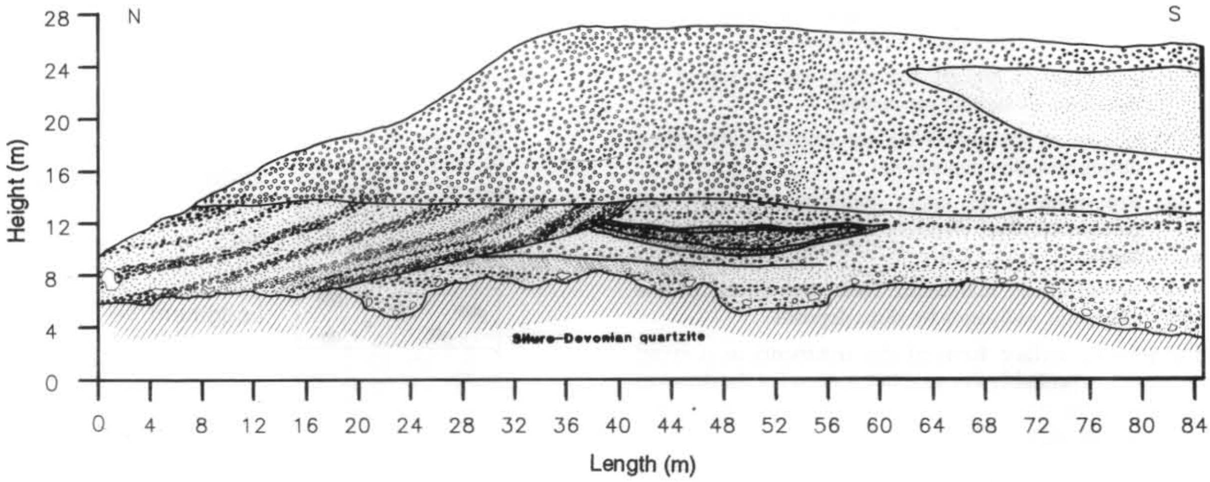


Figure 24. Type section of the Cableway Formation showing outwash gravel of a proximal outwash stream resting on Siluro-Devonian quartzite and overlain by locally derived slope deposits.

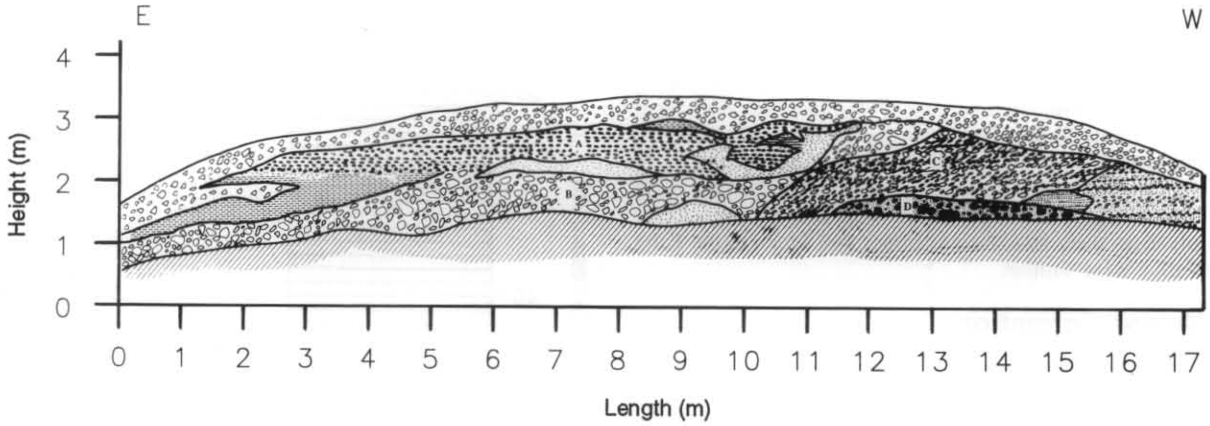


Figure 25. Ice-contact stratified sediments of the Cableway Formation resting on Thureau Formation gravel (filled pattern) and overlain by slope deposits. Letters correspond to fabric analyses shown in Figure 26.

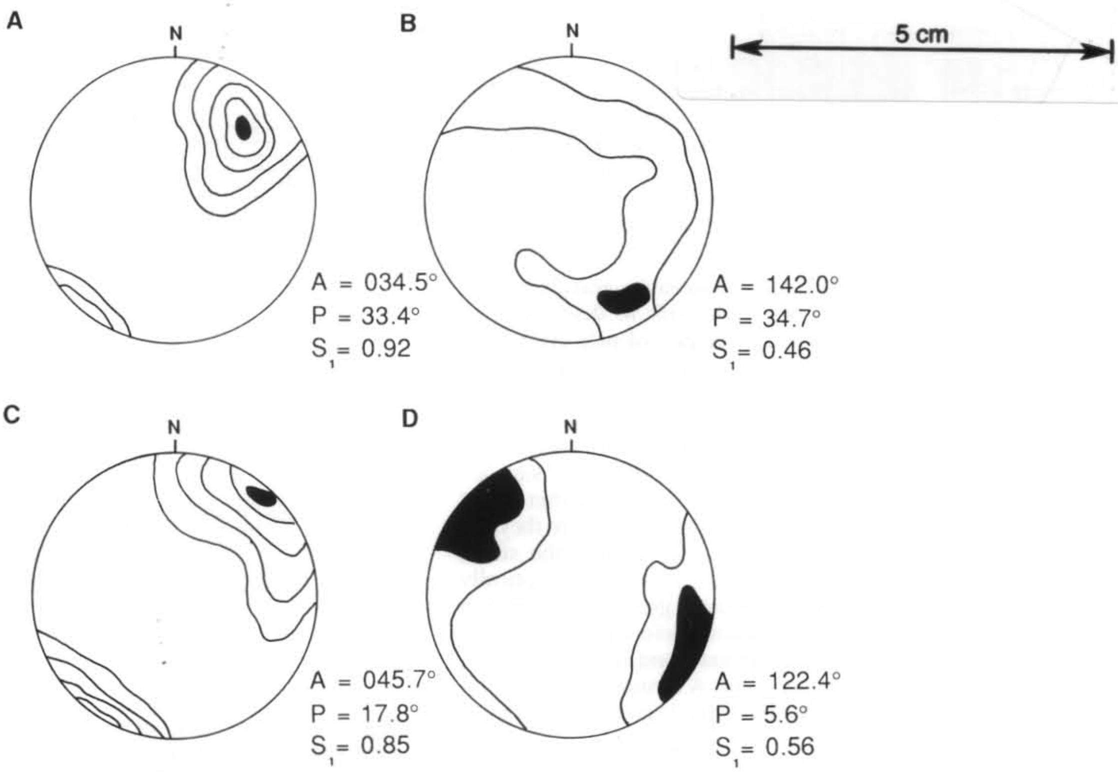


Figure 26. Pebble fabric of ice-contact sediments shown in Figure 25. The equal area nets are contoured at 2 standard deviation intervals using the method of Kamb (1959). The zone of maximum clustering is black.

(3) deposition of the massive diamicton on top of the sand, possibly by slumping from an ice surface;

(4) deposition of gravels and sandy silt on top of the diamicton as melting proceeded;

(5) melting of the underlying ice causing deformation during and after deposition;

(6) deposition of the upper undeformed gravels and paraglacial slope deposits as the buried ice finally melts;

(7) cessation of meltwater flow and burial of the glacigenic sediments by slope deposits.

The present surface form of the sediments as a small hummock is probably due to relief reversal of the deposits by accumulation in a basin on an ice surface similar to that described by Shaw (1972).

Although the limits of the Cableway and Thureau advances in the King Valley are similar (fig. 2), the evidence from the other distributary lobes clearly suggests that the Thureau advance was much more extensive. This apparent anomaly is probably due to the nature of the glacial system in which the watersheds of Comstock and Linda Creek controlled the thickness of ice in the King Valley. During the Thureau advance the Comstock Gap was breached by ice that flowed into the Queen Valley and the ice terminated just east of Karlsons Gap at the head of the Linda Valley. However, during advances when ice thicknesses were not sufficient to breach the divides or fill the tributary valleys the ice was concentrated in the King Valley. Thus, although the ice volumes of later glacier advances were less, their volumes and limits in the King Valley were similar.

Nelson Formation

The Nelson Formation is a 44 m sequence of laminated lake sediments that lie above the Cableway Formation and beneath the David Formation in the central part of the King Valley. The lake sediments rest on 10 m of gravel or till that is underlain by a layer of cobbles and Devonian siltstone (fig. 27). The sediments are known from drill cores and large excavations at CP896370, and a few small exposures in the banks of the King River. The formation has no surface expression.

The gravels at the base of the sequence are known only from drill cores and are not exposed. They consist of fine gravels with a maximum particle size of 100 mm and a mixed lithology that is dominated by locally derived Siluro-Devonian sediments. In some drill cores there appears to be a lag of large boulders at the base of the gravel just above the rockhead (J. Geidl, pers. comm., 1986; fig. 27).

The character of the contact between the gravels and the overlying laminated silts is difficult to determine but appears to be sharp. The silts consist of pale green beds with multiple sub-laminae less than one millimetre thick, and occasional dark grey laminae up to 40 mm in thickness (fig. 28A). According to the classification of Folk *et al.* (1970) the dark green laminae are silts and the pale green laminae are muds (fig. 28B). The silts are relatively uniform throughout the section to within 4 m of the contact with the overlying diamicton where they are intensely deformed. Just below the contact the silt is brecciated, and dark grey pieces of silt are chaotically distributed throughout the massive green matrix. Within 1.5 m of the brecciation the intense deformation gives way to gentle warping of the laminae, and numerous small, high-angle reverse faults occur. A further 2 m below this the unfaulted silts are gently warped and dip west at up to 17°.

The remainder of the silts have no other primary or secondary structures. Of particular note, and unusual for this kind of deposit in the King Valley, is the absence of

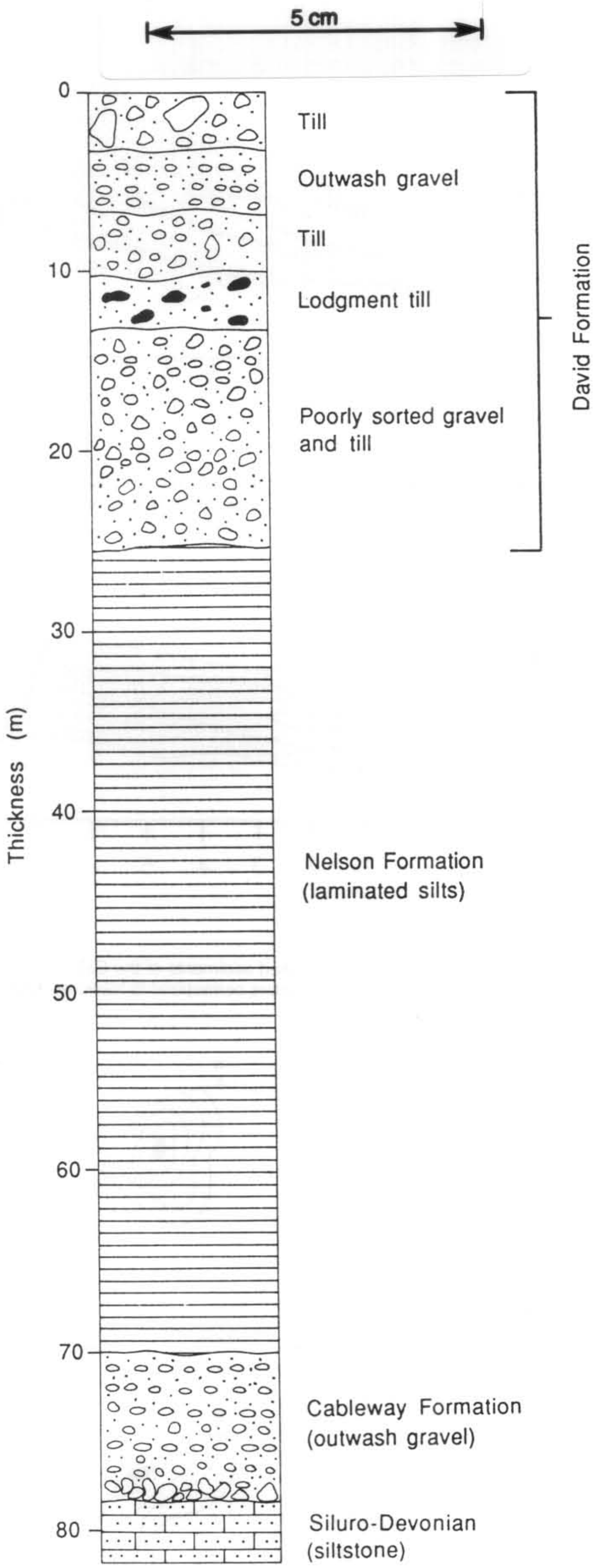


Figure 27. Type section of the David Formation showing the relationship between David Formation sediments (tills and outwash gravel), the Nelson Formation (laminated silts) and the Cableway Formation gravel which rest on Siluro-Devonian siltstone.

dropstones. Several samples were examined for fossil pollen but only three grains of Compositae were found, though the processed samples had traces of organic debris.

Numerous, flat, carbonate-cemented concretions occur in the dark grey silty laminae at around 190 and 210 m altitude. The origin of the carbonate is not known but similar concretions are common in Pleistocene glacial lake sediments.

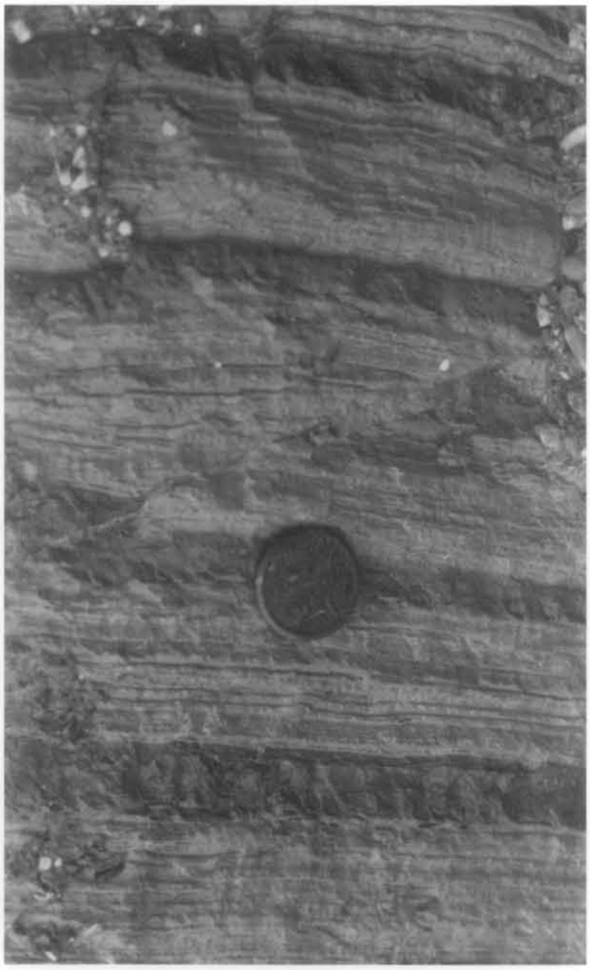
A seismic reflection survey of these unconsolidated sediments encountered several shallow-dipping reflectors within the silts (Robertson, 1986). Although Robertson interpreted the reflectors as possible evidence for sand lenses in the lake silts, none of the cores or excavations encountered such lenses. The reflectors may be related to the presence of horizons of carbonate concretions associated with the dark laminae.

Twenty-seven specimens were taken from exposures of the laminated silts of the Nelson Formation for palaeomagnetic analysis. These specimens were more strongly magnetised than those of Baxter Rivulet and Thureau Hills, NRM values ranging from 0.7 to 22.9 μG. Three quarters of the specimens had a fair capability of maintaining a stable remanence, whilst the remainder had a good capability. The specimens were demagnetised at 35 mT and all had a normal polarity (fig. 29). A further thirty-eight specimens from drill cores also had a normal polarity.

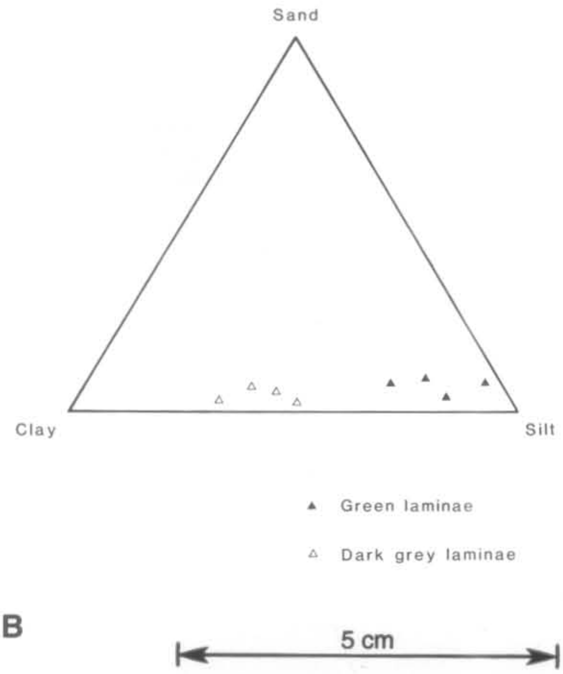
The laminated silts were probably deposited by suspension from the tails of turbidity currents in a deep lake. The temporal relationships of the individual laminae are not clear but investigations of sedimentation in temperate proglacial lakes suggests it is unwise to call such laminae varves and interpret the sedimentary couplet as the product of one years sedimentation (Smith *et al.*, 1982).

A lake of the size and depth necessary for the formation of sediments of this type must involve a large dam in the lower part of the King Valley. The most likely explanation for the damming is by a moraine deposited by the Cableway advance. The geomorphology of the valley suggests that this may have occurred about 1.8 km upstream of the Governor River confluence at the south-western edge of the Thureau Hills, where there is a constriction in the valley that could easily be blocked. The absence of pollen in the silts suggests that they either accumulated rapidly or in a cold environment or both. The thick, uninterrupted sequence of silt deposition, the absence of dropstones, scour, sediment flows and structures typical of ice-marginal lakes as described by Shaw (1975), suggests that the depositional environment was remote from direct glacial influences and without floating ice. The silts are therefore interpreted as representing an interstadial period separating the Cableway and David advances. The length of the interstadial can be estimated by assuming similar sedimentation rates to Holocene proglacial lakes in New Zealand. Pickrill and Irwin (1983) calculated the sedimentation rate in Lake Tekapo as 10 ± 1 mm/ yr^{-1}. Using this rate the 44 m of silts known from the David type section could have accumulated over a period of 4400 ±440 yr.

The interpretation is equivocal. An alternative explanation is that the silts accumulated when the climate was glacial and ice was confined to the upper part of the Tyndall-Lake Dora Plateau and the higher peaks of the West Coast Range. Regardless of climatic inferences, the silts clearly represent a significant temporal break in direct glacigenic deposition in the middle of the King Valley and lie between two glacial advances. On these grounds alone the deposit can be regarded as evidence for an interstadial.



A



B

Figure 28. Nelson Formation laminated silts and muds. A: A low angle reverse fault in laminated silts of the Nelson Formation. B: Ternary diagram of the particle size of the deposits shown in A.

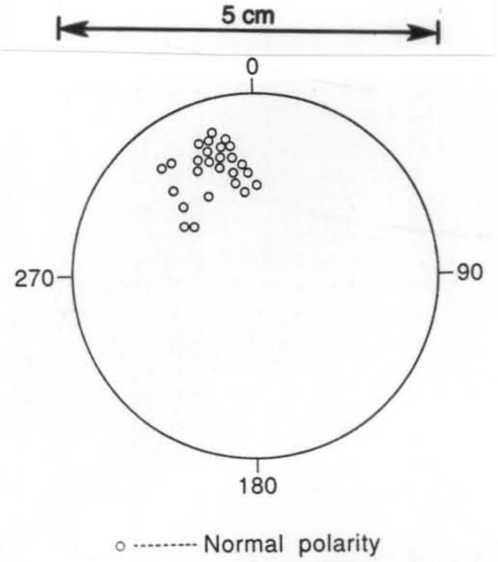


Figure 29. Stereoplot of the detrital remanent magnetisation of the Nelson Formation silts.

David Formation

The David Formation is the most extensive glacial formation in the King Valley (see Corbett *et al.*, 1989). Its recognition is based on two excavations and numerous drill holes through an end moraine and outwash surface at CP896370. The excavations show that the formation consists of 29 m of interbedded till and outwash gravel that overlie laminated silt of the Nelson Formation (fig. 27). In the lower King Valley near the Governor River numerous low terraces are inset into older Cableway Formation deposits and are interpreted as representing reworked gravels associated with outwash streams of the David Advance. Other large exposures of David Formation glacigenic sediments occur at the mouth of Linda Creek and near the Nelson River at Valley Creek.

The type section [CP896270] of the David Formation consists of two 3 m-deep trenches that run from a terminal moraine crest at 242 m to the King River at 194 m, and numerous drill holes that penetrate up to 80 m of Quaternary sediments resting on Siluro-Devonian siltstone (fig. 27).

The deposits exposed in the excavations form a sequence of sediment flows, outwash gravels, massive diamictons and lodgement till deposits. The sequence rests on the Nelson Formation laminated silts and muds and records the advance of the King Glacier and the formation of an end moraine. The near surface sediments have been affected by post-depositional mass movement. This has caused intense local plastic deformation, faulting and brecciation of the laminated silts, and formation of clastic dykes as fills of tension cracks in the overlying tills and outwash gravels (Fitzsimons and Colhoun, 1989). The contact between the silts of the Nelson Formation and the overlying diamicton is sharp and eroded in most places, but at two locations in the trenches pale green, well-sorted medium sands form what may be a transitional and conformable deposit. There appears to be a rapid change in the energy level of the environment, probably associated with the onset of the David ice advance.

The diamicton at the base of the David Formation (fig. 27) is massive, unsorted, and has a maximum particle size of 300 mm. The lithology consists of a mixture of locally derived West Coast Range and Eldon Range rocks. At 12 m and 14 m depth thin beds of laminated silt and crudely bedded gravel dip west at 17° (fig. 27). For the remainder of its thickness the deposit is massive but occasional small sand lenses occur.

The overlying 2.2 m-thick diamicton has a pronounced fissility, is highly consolidated, and consists of pebbles up to 50 mm in diameter supported in an iron-stained silty matrix. The diamicton overlying the highly consolidated diamicton is massive, has a maximum particle size of 300 mm. Where the diamicton is exposed in the trenches, several narrow wedge-shaped clastic dykes strike across the slope. The contact with the overlying outwash gravel is sharp and scoured.

The outwash gravel consists of horizontally bedded sands and gravels that dip west at 15°. The gravel is well sorted, has numerous voids many of which are lined by illuvial mud. Beds of sand within the gravel are coarse, poorly-sorted, massive and locally iron-cemented. The pebble fabric of the clasts suggests deposition by a current flowing toward 227°, obliquely away from the reconstructed ice margin.

The uppermost deposit in the sequence is a very coarse diamicton with boulders up to 1.7 m in diameter resting in a silty matrix (fig. 27). Numerous thin, gently warped silt stringers are draped over larger clasts, and some small flow noses are preserved. The matrix of the diamicton is a dull orange colour from iron-staining and has multiple iron pans.

The tills and outwash gravel deposits overlying the Nelson Formation record the advance of ice at least as far as, and possibly beyond the site, the *in-situ* melting of ice and the development of an end moraine. The diamictons overlying the silt and the consolidated diamicton have weak pebble fabrics, unrelated to ice flow direction. They are thought to have been deposited as sediment flows in a supraglacial depositional environment. The highly consolidated nature, strong fissile structure and strong unimodal pebble fabric of the overlying diamicton suggest it is a lodgement till. Because lodgement tills are always the basal sediments of an advance (Boulton, 1976), the presence of this till is evidence that the glacier advanced beyond this site during the David advance.

The presence of large sub-angular erratic boulders up to 1.7 m in diameter, low compaction, pebble fabric and the washed character of many parts of the uppermost diamicton suggest that it was deposited by mass movement. Such movement probably included several small sediment flows and slumping from a supraglacial till on a melting ice surface. This is supported by the surface topography of the arcuate moraine ridge containing the supraglacial till which appears to have accumulated as a latero-frontal apron of debris adjacent to a melting ice edge.

The reconstructed sequence of events that produced these sediments is:

(1) damming of the lower King Valley by a moraine ridge during the retreat of the King Glacier from the maximum position of the Cableway advance;

(2) formation of the thick deposit of laminated silts (Nelson Formation) as bottom sediments in a large nonglacial or proglacial lake;

(3) advance of the King Glacier, erosion of parts of the upper sequence of lake sediments and deposition of massive sediment flows interbedded with outwash gravels at the front of the glacier;

(4) deposition of lodgement till as the ice advances into the valley;

(5) retreat of the glacier to a position near the section with deposition of outwash gravels by a proglacial stream, followed by the formation of sediment flows, slump deposits and laminated silts that were formed at a relatively stable ice front. These sediments formed an arcuate latero-frontal moraine ridge;

(6) post depositional deformation of the silts is associated with the formation of tension cracks and landslides.

In addition to the main outwash plain, several terraces are cut into older deposits downstream of the type section. On

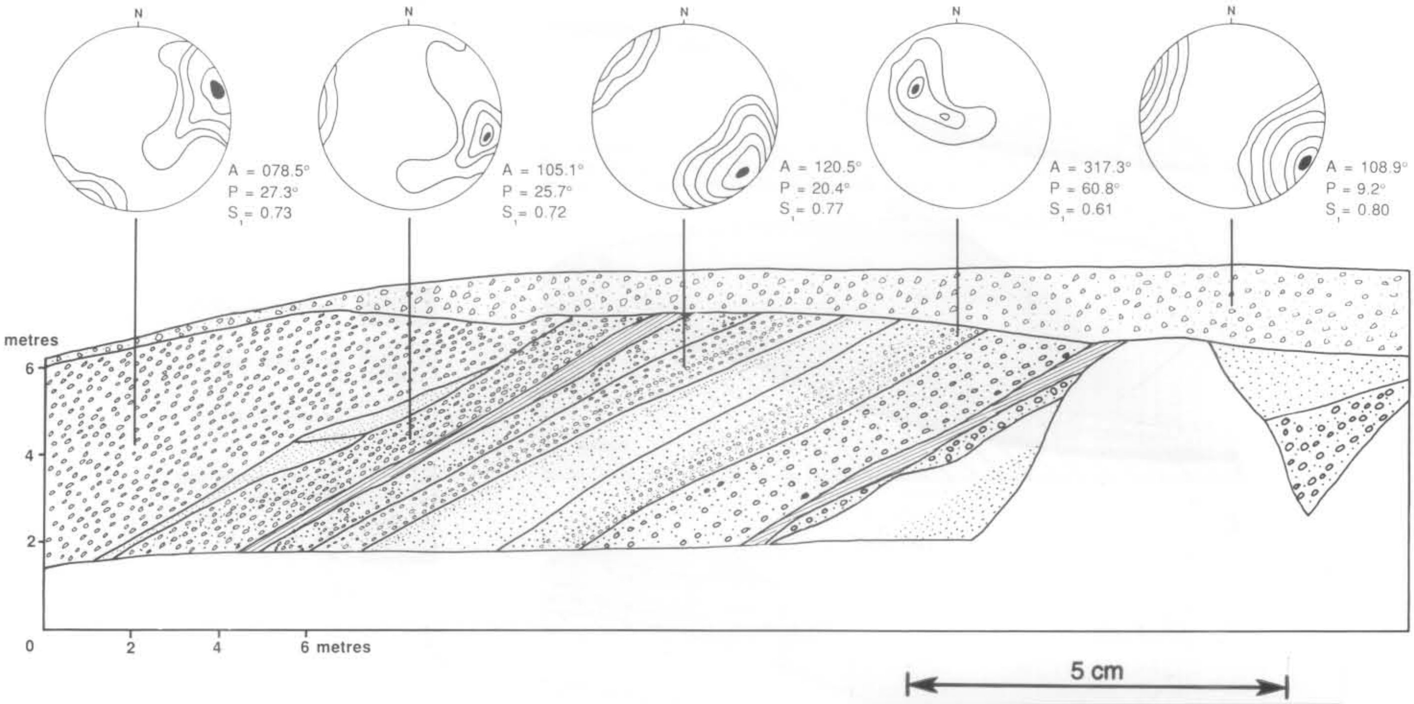


Figure 30. Gilbert-type delta exposed at the mouth of the Linda Valley showing steeply-dipping foreset beds. The equal area nets are contoured at 2 standard deviation intervals using the method of Kamb (1959). The zone of maximum clustering is black.

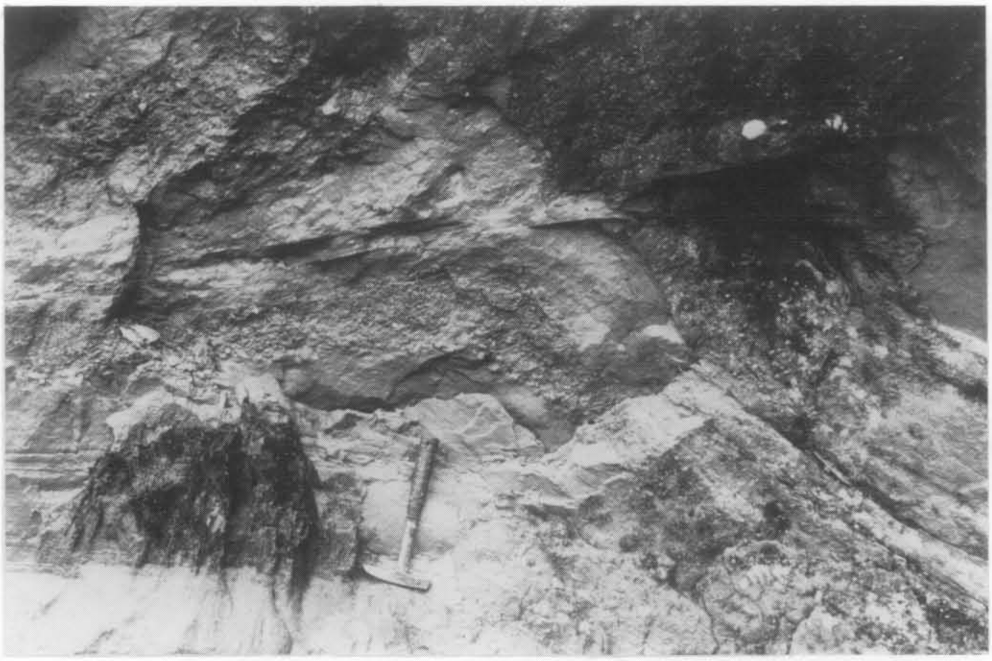


Figure 31. Deformed laminated sediments exposed at the mouth of Linda Creek. Note the low-angle shear plane overlain by lodgement till indicating that the sediments have been overridden by glacier ice.

the Crotty Plain at the confluence of the King and Governor Rivers substantial erosion and redeposition of the Thureau, Moore and Cableway formations took place during the David advance.

At the mouth of Linda Creek there is a series of sections through glacial sediments at the same altitude as the type section of the David Formation. These sediments were deposited as ice-contact terraces and lacustrine deposits as the King Glacier blocked Linda Creek. They record a series of ice-marginal lacustrine environments dominated by high energy subaqueous debris flows.

The first of these sections is at CP872409 and shows a series of interbedded gravels, sands and laminated silts that dip east into the King Valley at up to 28° (Fitzsimons, 1988) (fig. 30, 31). The sediments record a delta deposited in a small ice-marginal lake that was fed by stagnant melting ice that formed further up the valley at an earlier stage of the ice advance. The delta prograded from west to east, that is toward the ice dam and not from the ice margin. The source of debris is unclear but the high altitude of the deposits relative to the valley floor suggests that the source was an ice mass that remained in the Linda Valley long after the initial glacial retreat.

The predominant mode of transport on delta foreset slopes is by mass flow and avalanching (Postma and Roep, 1985), and is dominated by bedded gravels and sands

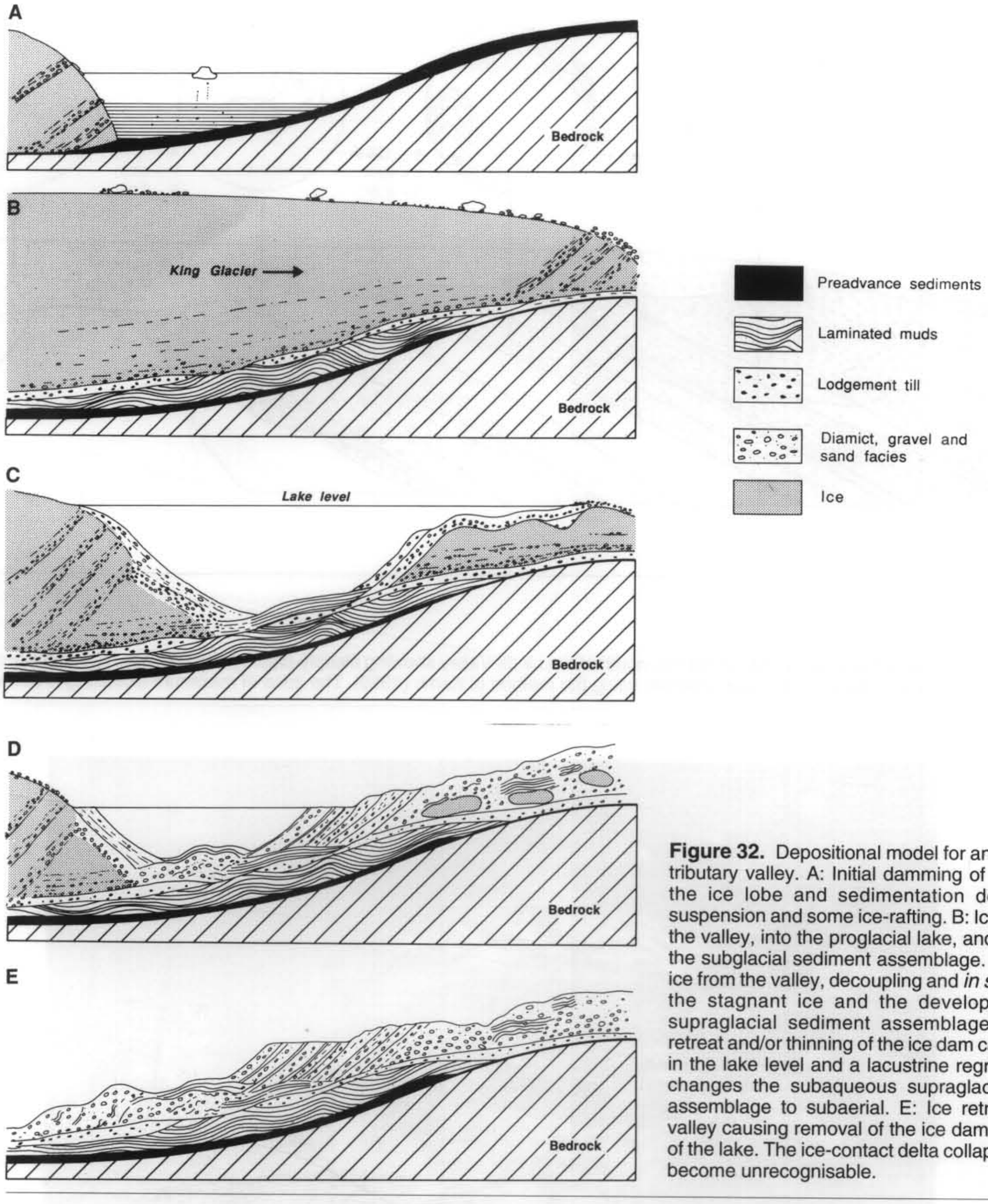


Figure 32. Depositional model for an ice-dammed tributary valley. A: Initial damming of the valley by the ice lobe and sedimentation dominated by suspension and some ice-rafting. B: Ice advance up the valley, into the proglacial lake, and formation of the subglacial sediment assemblage. C: Retreat of ice from the valley, decoupling and *in situ* melting of the stagnant ice and the development of the supraglacial sediment assemblage. D: Further retreat and/or thinning of the ice dam causing a drop in the lake level and a lacustrine regression which changes the subaqueous supraglacial sediment assemblage to subaerial. E: Ice retreat from the valley causing removal of the ice dam and draining of the lake. The ice-contact delta collapses and may become unrecognisable.

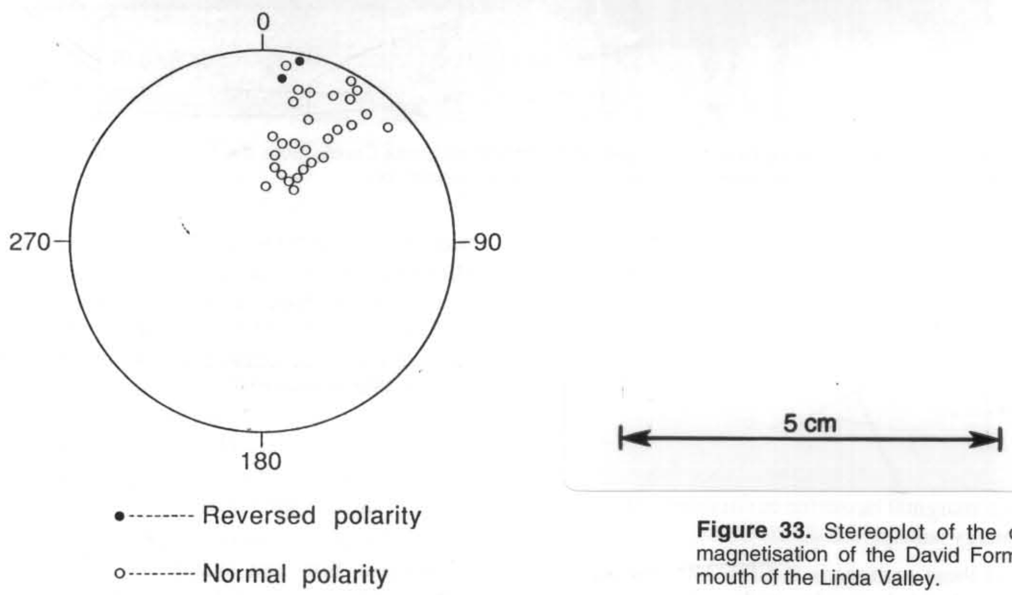


Figure 33. Stereoplot of the detrital remanent magnetisation of the David Formation silts at the mouth of the Linda Valley.

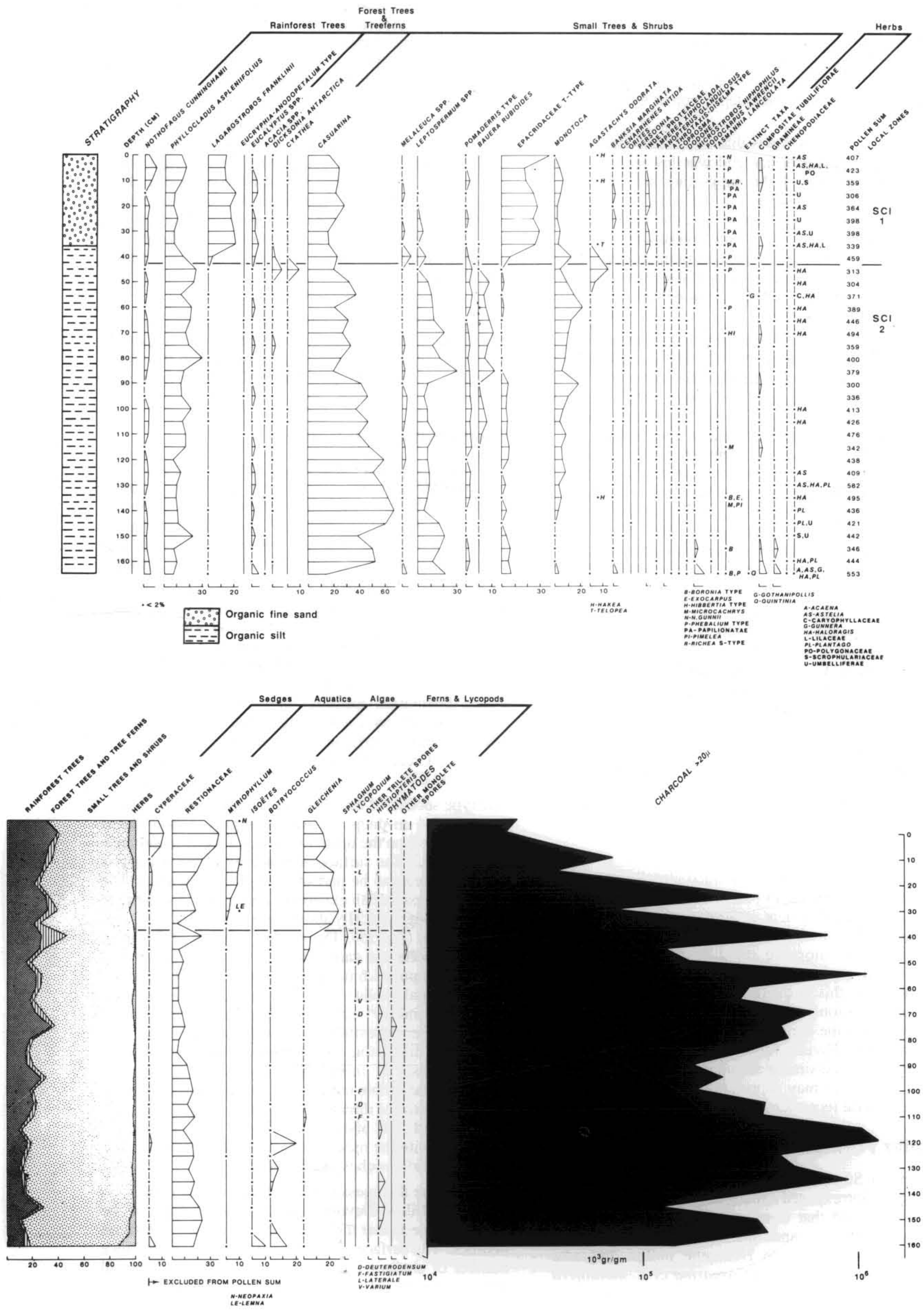


Figure 34. Smelter Creek Pollen diagram

often with multiple scour surfaces (Cohen, 1979). The multiple occurrences of laminated silts seen in this section are not typical of delta foreset environments, and are interpreted as being draped over sliding and avalanching gravels during quiescent periods. Dropstones within the laminated sediments suggest that there was floating ice on the surface of the lake during deposition of the laminated sediments.

An alternative explanation is that the whole sequence of deposits accumulated on ice and was lowered into this position during melt of the supporting ice. However, because the pebble fabric of the gravels suggests deposition on a steep slope and because the sediments lack the intense deformation that would accompany lowering, the sediments are thought to have accumulated on a delta slope. Figure 32 is a reconstruction of the environment of deposition, an ice marginal lake with a delta prograding into it from Linda Creek.

About 100 m west of the section described above at CP870409, a large section 300 m long and 15 m-high exposes a sequence of deposits that have a similar geometry to the section described above. Although most of the section is covered by moss, occasional clearings show diamictons interbedded with steeply dipping silts. The sediments appear to have accumulated on a slope that has the same geometry as the delta deposits described above. The eastern edge of the section shows a sequence of laminated silts underlain by gravels and overlain by lodgement tills and slope deposits (fig. 31). The laminated silts lie below the series of subaqueous sediment flows that are poorly exposed in the western extension of the section.

Thirty-one specimens for palaeomagnetic analysis were taken from laminated silty sands, sandy silts and silts. These specimens showed a similar magnitude of magnetisation to the Nelson Formation, with NRM values ranging from 1.0 to 21.8 μG. The majority of the specimens had a fair capability of maintaining a stable remanence, with a few specimens having a good capability. Specimens were demagnetised at 25 mT after step-wise demagnetisation of representative specimens. Twenty-nine of the thirty-one specimens had a normal polarity (fig. 33).

Bull Rivulet Formation

The Bull Rivulet Formation is named after its proximity to Bull Rivulet in the upper part of the valley. There is no type section because there is no exposure of the small moraine or outwash surface remnant. However, a shallow pit dug on the moraine crest exposed weathered, poorly-sorted gravels of glacial origin. The geographic distribution of this formation is very limited because subsequent erosion has destroyed most of the outwash plain. Although the duration of the period separating the David and Bull Rivulet formations cannot be assessed, their proximity and similar altitudes suggest that the Bull Rivulet sediments may represent a minor ice advance during a general recession from the David advance.

Smelter Formation

Adjacent to Smelter Creek, near the old township of Crotty, a section exposed a sequence of sands, gravels and organic fine sand that were analysed for pollen. Pollen from the organic fine sands and silts shows a very similar vegetation assemblage to Holocene pollen diagrams except for the abundance of *Casuarina* cf *C. monilifera* pollen (fig. 34). A lower pollen zone SC2 can be identified by substantial quantities of *Phyllocladus and Casuarina* that are accompanied by the wet forest/scrub taxa *Pomaderris*, *Bauera*, *Monotoca* and *Leptospermum*. The absence of taxa that are normally found at higher altitudes today suggests that the vegetation was lowland scrub-rainforest with abundant *Casuarina*. This may indicate a pioneer stage of development more like the early Holocene rather than the middle Holocene maximum development of rainforest. Wood from the deposit gave an amino acid age that suggests a minimum age equivalent to Isotope Stage 5 (B. J. Pillans, pers. comm., 1988).

Above 400 mm (SC1) two notable changes occur. These include the strong expansion of *Lagarostrobos* and of Epacridaceae T-type. Both suggest a trend toward wetter conditions. *Lagarostrobos* would have expanded along the river valleys while Epacridaceae would have been important in the wet heath on the poorly-drained higher sites. In addition, areas of scrub-rainforest, probably also occurred on higher sites with moderate to good drainage. The absence of significant herbs indicates that at all times it was a lowland-scrub-heath environment.

A single spore of *Gothanipollis perplexus* and one grain of *Quintinia psilatispora* were recorded. If the suggested minimum age for this deposit is correct then this is the youngest record for these normally ancient taxa in Tasmania.

Chamouni Formation.

Outwash gravels of the Chamouni Formation form extensive terraces inset into older glacial deposits (Corbett *et al.*, 1989). The upstream extent of these terraces is unclear but they seem to be buried by an alluvial fan at the confluence of the Eldon and South Eldon rivers. The formation is named after the Chamouni Valley, which is an old name for the Linda Valley.

The Chamouni Formation is limited in extent to the upper King Valley near the Lyell Highway and adjacent to the mouth of the Linda Valley. It consists of a series of terraces inset in older glacial sediments that can be traced 4 km downstream of the type section to a position near the David Formation terminal moraine. Although most exposures of the terrace consist of bedded outwash gravels unconformably overlying laminated silt, an exposure near Comstock Creek shows that part of the terrace consists of lodgement till.

The type section of the Chamouni Formation consists of a small 2.4 m-deep section of highly consolidated lodgement till on the upstream end of a roche moutonnée at CP896432. The surface morphology of the outcrop is flat and appears to be part of a subglacial till sheet. The sheet forms part of an extensive terrace at 236 m altitude which is topographically above the lower part of the Dante outwash fan (fig. 35). The section consists of 2.4 m of highly consolidated dark grey silty till overlain by a thin soil. In the upper 1.6 m the structure of the till exhibits a sub horizontal fissility. The fissility dips to the north at between 5 and 10°, and appears to be due to shearing during or after deposition. There is no textural variability within the diamicton which consists of subangular to rounded clasts of mixed lithology up to 150 mm in diameter. The pebble fabric of the diamicton is strong and unimodal with the direction of maximum clustering (V_1) being toward 151° and dipping at 17.8° (fig. 37). This is consistent with the reconstructed ice flow direction from striae on nearby roches moutonnées.

Although the thickness of the sediment is not known, an outcrop of Siluro-Devonian siltstone 60 m south of the section suggests that the till is not very thick and rests on bedrock. The surface of the terrace has been eroded by a number of shallow channels and the surface of the till shows no signs of a lag deposit and appears to be a depositional surface.

The highly consolidated nature of the diamicton together with the fabric, fissility, texture and its location on the upstream side of a roche moutonnée, suggest that it is a lodgement till. The only apparent inconsistency in the

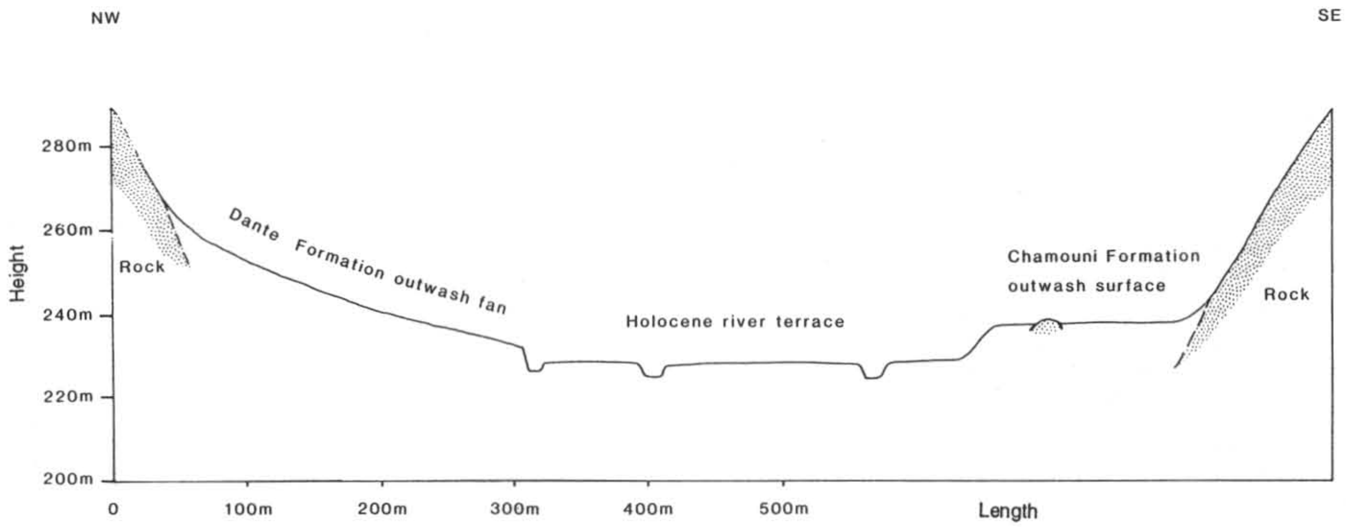


Figure 35. Topographic profile showing the relationship between the aggradation surfaces of the Chamouni and Dante formations.

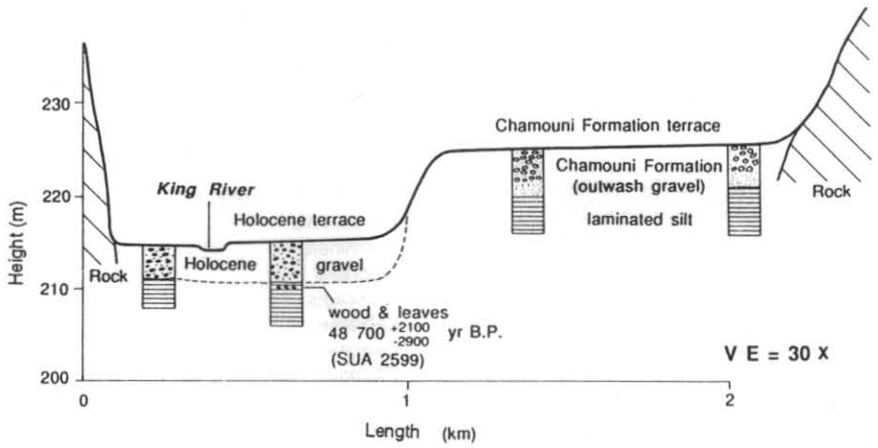


Figure 36. Chamouni Formation sediments south of the type section showing the relationship between the Chamouni Formation and Holocene alluvial gravels.

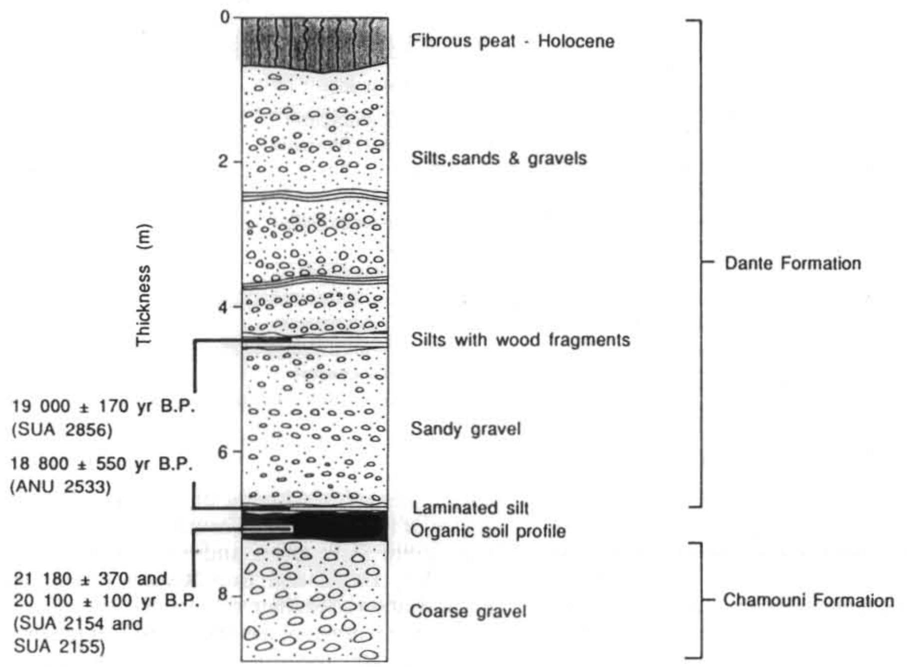


Figure 37. Type section of the Dante Formation (From Gibson *et al.*, 1987 with modifications).

5 cm

interpretation is that the pebble fabric dips down glacier whereas other described pebble fabrics of lodgement tills dip up glacier (Boulton, 1976). This may be due to local stresses and flow deflection on the upstream side of a rock obstruction.

Weathering rinds on Jurassic dolerite have a mean thickness of 1.53 mm and a standard deviation of 0.7 mm, which is very similar to that described by Kiernan (1983*a*) for the Dante Formation. The matrix is unweathered apart from some minor bleaching and iron-staining which is restricted to the top 200 mm of the soil profile.

South of the type section several small roches moutonnées rise above the terrace surface, and numerous exposures of outwash gravel rest on organic-rich sands and silts. About 800 m south of the type section near the King River Road at CP889427 an exposure in the Chamouni Formation terrace shows an upward coarsening sequence from organic silts to sand, coarse sand, pebbly sand and gravel (fig. 36).

Although the depth of the dark grey silts is unknown, auger holes show that they are at least 4.5 m thick. The silts are very finely laminated, with individual laminae from less than 1 to 4 mm thick. Abundant small drifted wood fragments, all less than 2 mm in diameter, occur in the uppermost 4 m of the silt which smelt sulphurous and appeared to contain a high organic content. Five samples at various depths were processed to extract pollen but only five pollen grains of Compositae were observed in all the samples. The silt grades upwards into massive medium silty sands and fine pebble gravels.

The upward coarsening sequence seen in the section suggests that the gravels are conformable on the silts. The sequence records the changing depositional conditions with the onset of the Chamouni ice advance. Weathering rinds on Jurassic dolerite in these gravels and on the till seen at the type section suggest they are of similar age to the Dante Formation and much younger than the David Formation.

An excavation in the Holocene alluvial surface near the King River Road at CP885422 shows low-lying Holocene gravels unconformably overlying the grey silts of the Chamouni Formation (fig 36). The lowermost deposits in the excavation consist of dark grey silts that are finely laminated and apparently identical to silts that crop out in the section described above. In this section the silts grade into sands containing organic debris including drifted twigs and leaves.

Analysis of pollen from the silts shows low quantities of trees and woody taxa and high quantities of herbs (table 6). The abundance of Gramineae, Compositae Tubuliflorae and Umbelliferae together with high Epacridaceae T-types and Cyperaceae suggests a non-forest herbland with sedges and heather plants. The occurrence of small quantities of alpine-subalpine shrub pollen together with *Donatia*, *Plantago* and *Oreomyrrhis* point to the vegetation being alpine to subalpine. This suggestion supports the interpretation that the transition from the silts to the outwash gravels represents the onset of a glacial advance (table 6). The wood was dated at 48 700^{+2900}_{-2100} yr B.P. (SUA 2599) and is overlain by low-lying Holocene river gravels (fig. 36). Since the date is at the limits of ^{14}C dating it must be regarded as a minimum age.

Excavations in the southern most extension of the Chamouni Formation terrace, 3.2 km south of the section described above at CP885404, show a similar pattern of outwash gravels overlying laminated silts. In these sections the silts grade upwards into massive sandy silts, sands and pebbly sand over a vertical interval of about 600 mm. At about 20 mm below the contact with the overlying gravels the sand contained fragments of drifted wood and leaves. These were not dated but are believed to correlate with dated wood and leaves in a similar position upstream (fig. 36).

Table 6. POLLEN FROM THE CHAMOUNI FORMATION

Taxa	*Per Cent*	
Temperate rainforest trees	**6.6**	
Lagarostrobos franklinii		2.9
Nothofagus cunninghamii		2.9
Phyllocladus aspleniifolius		0.8
Other trees and ferns	**5.6**	
Dicksonia antarctica		0.3
Eucalyptus		5.3
Small trees and shrubs	**12.1**	
Casuarina		0.5
Coprosma		1.3
Epacridaceae T-type		7.5
Leptospermum		0.5
Orites		0.5
Papilionatae		0.5
Telopea		0.3
Tasmannia lanceolata		0.5
Pimelea		0.5
Alpine-Subalpine trees shrubs	**3.6**	
Athrotaxis-Diselma archeri type		0.3
Microcachrys tetragona		0.3
Microstrobos niphophilus		2.7
Nothofagus gunnii		0.3
Herbs	**71.7**	
Chenopodiaceae		0.3
Compositae Tubuliflorae		14.5
Cruciferae		0.3
Donatia novae-zelandiae		1.0
Gentianella		0.5
Gramineae		18.2
Gunnera		1.3
Haloragis		2.7
Labiatae		0.3
Liliaceae		0.3
Plantago		2.1
Polygonaceae		0.5
Ranunculus		4.5
Scrophulariaceae		0.5
Umbelliferae		20.1
Hydrocotyle		0.3
Oreomyrrhis		4.3
Total pollen and spores in sum = 373	**99.6**	
OUTSIDE PERCENTAGE SUM		
Sedges	**8.5**	
Cyperaceae		7.5
Restionaceae		1.0
Aquatics	**0.8**	
Isoetes		0.3
Myriophyllum		0.5
Fern spores	**2.9**	
Gleichenia		0.5
Lycopodium fastigiatum		0.3
L. scariosum		0.3
Other trilete spores		0.8
Histiopteris		0.3
Other monolete spores		1.0

The degree of weathering of both the till and outwash gravel of the Chamouni Formation are similar to that of the Dante Formation and very different from the preceding David and Bull Rivulet formations. These considerations, together with the location of the deposits between the limits of the Dante and David advances and the minimum age from the ^{14}C date suggest it was deposited during the early part of the Margaret Glacial Stage (table 1).

Dante Formation

The Dante Formation is named after Dante Rivulet which flows from the Lake Dora Plateau into the King River. Figure 35 shows the small, steep Dante outwash fan and its relationship to the Chamouni Formation outwash surface. Unfortunately, no section showed a contact between the Dante and Chamouni formations. A similar though larger fan with an end moraine at the apex occurs in the South Eldon River near its confluence with the King River.

The type section of the Dante Formation at CP902456, was described by Kiernan (1980). It consists of outwash gravel overlain by a palaeosol that in turn underlies outwash sediments of the Dante advance (fig. 37). Wood 100 mm from the base of the Dante outwash was dated at 18 800 ± 550 yr B.P. (ANU 2533). Pollen from the palaeosol records an alpine herbfield-bog mosaic and contains a macrofossil of the alpine cushion plant *Donatia novae-zelandiae* (Gibson *et al.*, 1987). The *Donatia* was dated at 21 180 ± 370 yr B.P. (SUA 2154) and drifted twigs stratigraphically below the *Donatia* were dated at 20 100 ± 470 yr B.P. (SUA 2155) (Gibson *et al.* 1987). Although Kiernan (1980) and Gibson *et al.* (1987) describe the lower outwash gravel as part of the Comstock Glaciation (= Henty Glaciation) it is probably part of the Chamouni Formation described above and belongs to the Margaret Glacial Stage.

An additional date from the site was obtained from wood at 4.35–4.60 m in 1989. The wood was dated at 19 100±1 70 yr B.P. (SUA 2856) and indicates that most of the fan accumulated quickly.

The gravels at the base of the section consist of well-sorted, massive imbricated gravels with a mixed lithology. They are separated from the overlying gravels by a lens of organic sandy silt which forms part of a weakly developed alpine palaeosol. The palaeosol is marked by roots and vertical plant stems, suggesting that the sandy silt was deposited in a wet, boggy environment, possibly in standing water. Pollen from the silt records an alpine herbfield-bog mosaic with a rich flora of herbs, sedges and low shrubs (Kiernan, 1980; Gibson *et al.*, 1987).

The upper outwash unit consists of 4.5 m of moderately well-sorted horizontally bedded outwash gravel, sands and silts. Weathering rinds on Jurassic dolerite clasts have a mean thickness of 1.5 mm and a standard deviation of 0.27 mm. The lithology of the pebbles is a mixture of West Coast Range and locally derived rocks.

Kiernan (1980) interpreted the section as demonstrating the formation of a soil during a period when the treeline remained below 230 m and considered that the grey sandy silt represented renewed glacial sedimentation after a break during which the palaeosol had formed (*ibid.*, p. 35). Gibson *et al.* (1987) concluded that the wood dated at 18 800 ± 500 yr B.P. (ANU 2533), a fossil bolster plant dated at 21 180 ± 370 yr B.P. (SUA 2154) and twigs from the palaeosol dated at 20 100 ± 470 yr B.P. (SUA 2155) imply that the overlying outwash gravels represent the late last glacial maximum. The Dante Rivulet section has subsequently become the unofficial type site for the maximum of the Last Glaciation in Tasmania (Colhoun, 1985*a*, 1985*b*; Colhoun and van de Geer, 1986).

Long Marsh Formation

Holocene sediments in the King Valley consist primarily of low-lying river terraces that are mainly limited to the upper part of the King Valley near the Lyell Highway. The terraces occur at two levels: the lower ones are within 1.5 m of base flow river level and are periodically flooded, while the higher ones are 3.5–4 m above the river level and are not subject to flooding. The deposits consist of about 1.5 m of buff-coloured, overbank, silty sands resting on up to 3 m of fluvial gravels which have been reworked from older glacial sediments which they overlie. Smaller areas of fluvially reworked gravels occur also in the Nelson River and in the lower King Valley.

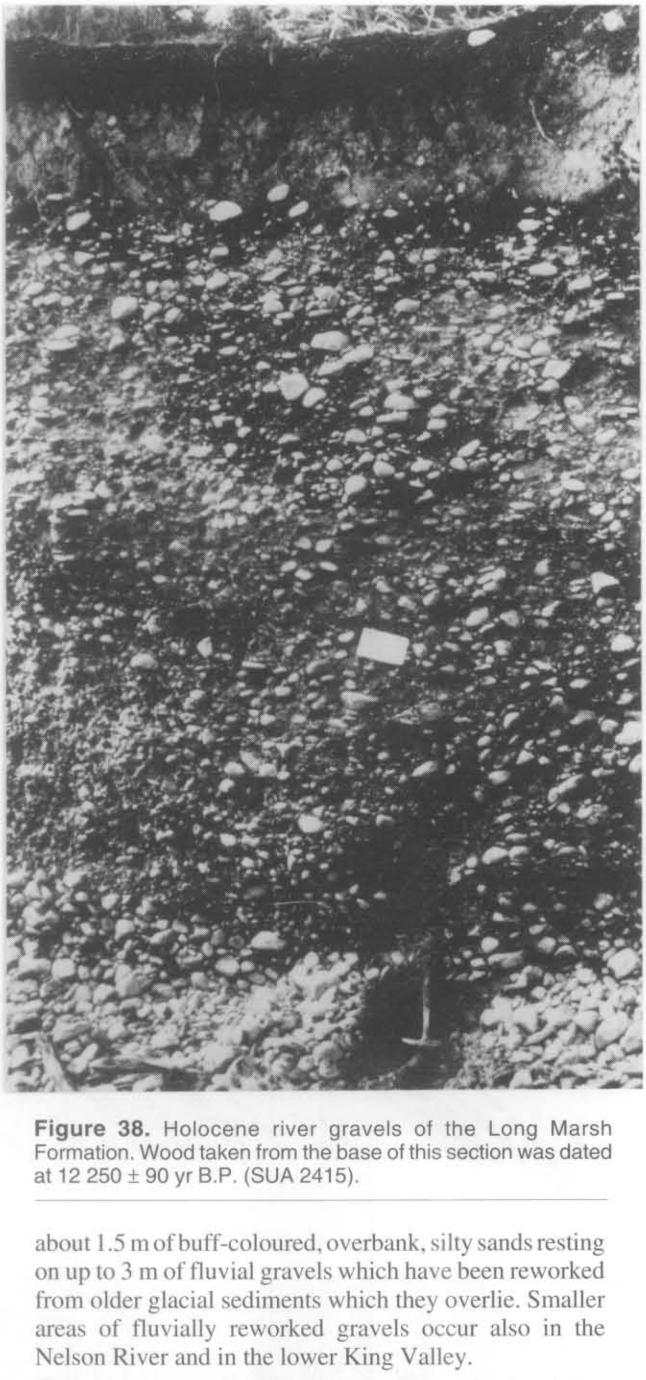


Figure 38. Holocene river gravels of the Long Marsh Formation. Wood taken from the base of this section was dated at 12 250 ± 90 yr B.P. (SUA 2415).

The type section of the Long Marsh Formation is a 3.5 m exposure in a terrace on the left bank of the King River at CP892378 (fig. 38). The gravel is moderately well-sorted, crudely bedded and has a strong imbrication which suggests it was deposited by a current flowing from north-west to south-east. Below 1.6 m the gravel is diffusely iron stained and cemented. The weathering rinds on Jurassic dolerite clasts have a mean thickness of 0.42 mm and a standard deviation of 0.16 mm.

At 3.5 m depth large logs up to 400 mm in diameter are buried in the gravel (fig. 38). A core sample of a large log was ^{14}C dated at 12 250 ± 90 yr B.P. (SUA 2415). This date suggests that large trees were growing in the King Valley prior to 12 000 yr B.P. and that the upper King Valley was ice free before this time.

About 100 m north-west of the type section, excavations in the terrace show Holocene gravels resting unconformably on grey silt. The silt is the same sediment that is exposed below the Chamouni Formation terrace (fig. 36). The King River level is now 3.5 m below the surface of the terrace, and a meander is eroding the bank

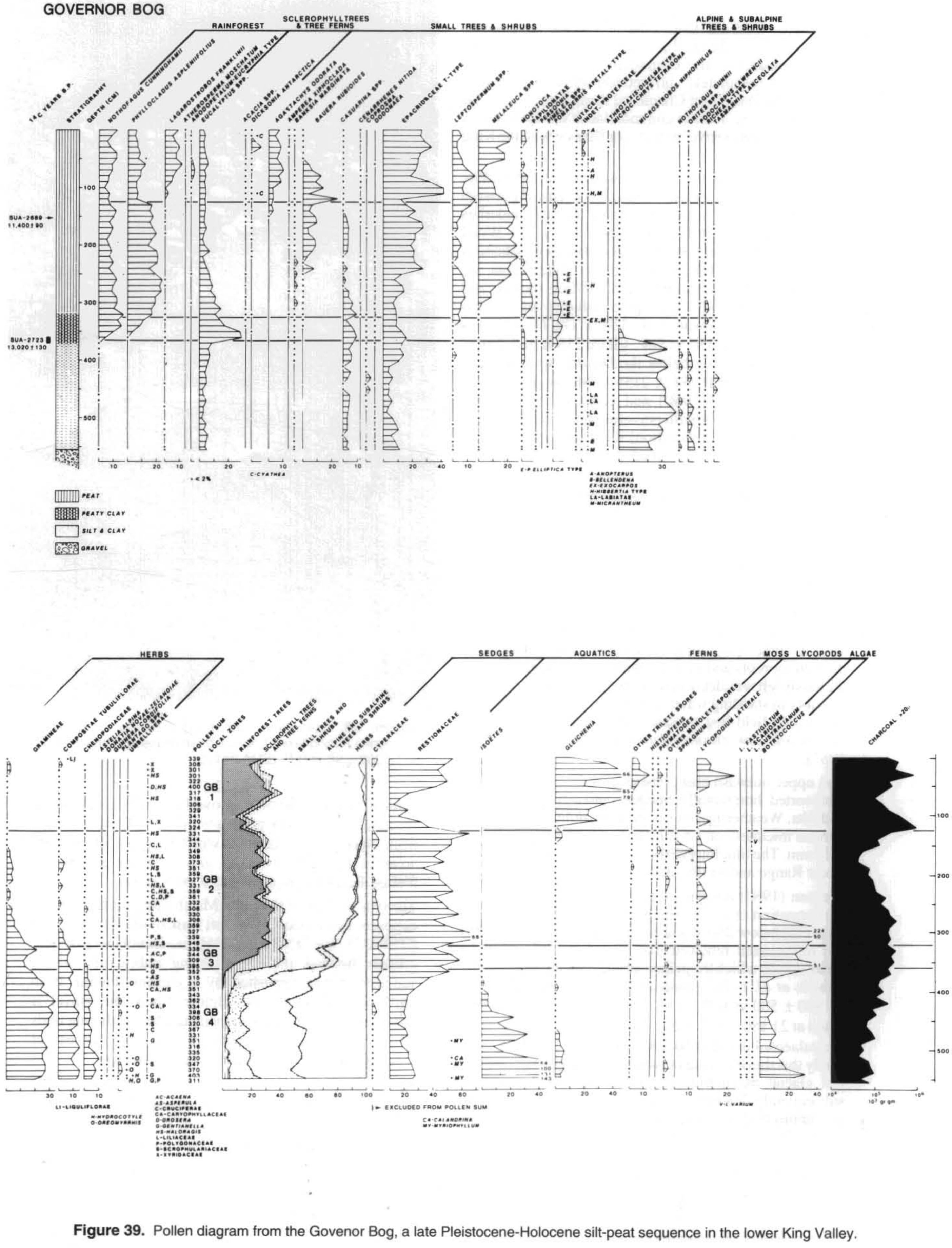


Figure 39. Pollen diagram from the Govenor Bog, a late Pleistocene-Holocene silt-peat sequence in the lower King Valley.

5 cm

and depositing a point bar 2 m below the level of the terrace.

The most common non-alluvial Holocene sediments are peats which form a blanket over most surfaces with less than 20° slopes. A trench and core from a doline near the Governor River has provided a good record of late glacial and Holocene sediment and vegetation changes for the King Valley.

The late glacial deposits consist of grey silt and mud that appear to rest on glacial gravels. These deposits pass upwards through peaty clay dated at 13 010 ± 130 yr B.P. (SUA 2723) to Holocene peat with abundant Restionaceae roots.

Pollen analysis of the sediments has resulted in the recognition of four pollen zones that correspond with changes in the vegetation (fig. 39). Three coincide with the sedimentary units while GB1 occurs in the Holocene peat.

Pollen zone GB4 extends from the base to 3.7 m depth. It coincides with the deposition of late glacial silt and clay before 13 010 ± 130 yr B.P. The major floral components are the subalpine shrub *Microstrobos niphophilus*, and Gramineae and Compositae. In addition, there are relatively high values for *Nothofagus gunnii*, *Orites* sp. and Chenopodiaceae which is typical for Tasmanian late glacial assemblages. The abundance of *Isoetes* indicates the presence of a shallow water oligotrophic lake with a mineral sediment bottom. The vegetation was alpine or subalpine herbland or shrubland during this period and rainforest was not present. The only trees present were *Eucalyptus* and probably *Casuarina* which can extend into the subalpine zone today. The climate was cool and wet.

Pollen zone GB3 occurs between 3.65 and 3.25 m, and represents a transition during which forest development commenced. The increasing quantity of vegetation and decrease of bare soil areas is reflected by the increased organic component of the sediment. The zone is characterised by a strong rise and fall of *Eucalyptus* while the two main rainforest taxa *Nothofagus* and *Phyllocladus* expand. The expansion of *Eucalyptus* prior to the full development of rainforest closely parallels the pattern at Tullabardine Dam further to the north in the Mackintosh Valley (Colhoun and van de Geer, 1986).

Relatively high values for *Pomaderris apetala* and *Monotoca* occur and indicate that the climate had become wet and warm. The *Pomaderris* bulge is a persistent feature in many early to middle Holocene pollen diagrams from the rainforest and wet sclerophyll forests of south-eastern Australia (Macphail, 1983). At the top of this zone Gramineae and Compositae begin to decrease to much lower values as rainforest expands. The wet conditions are also indicated by the increase in Cyperaceae and Restionaceae, and by the maintenance of the small pool in which *Botryococcus* flourished.

Pollen zone GB2 from 3.25 to 1.25 m represents the major period of Holocene temperate rainforest development. While *Phyllocladus* and *Nothofagus* are well represented, wet scrub and heath were also present around the site as indicated by the high values for *Melaleuca*, *Bauera rubioides* and Epacridaceae. The shallow lake became overgrown with Cyperaceae and Restionaceae and ceased to be a lake above 2.8 m. At this time *Botryococcus* disappears and *Melaleuca* reaches its maximum development. Abundant wood of *Melaleuca* in the peat indicates that it grew on the swamp surface.

Pollen zone GB1 from 1.25 to 0 m still represents cool temperate rainforest, wet scrub and heath vegetation but there are minor changes in the vegetation. There is a strong development of *Lagarostrobos* which would probably have migrated slowly along the river banks. There is a strong expansion of *Agastachys odorata*, a proteaceous shrub that occurs in openings in wet forest and in scrub and heath communities. *Gleichenia* also expands to high values and probably was widely distributed in the riparian and heath vegetation. Although the vegetation of this region has been altered by fire associated with mining and logging activities in the late 19th century (Kirkpatrick, 1977*b*), the alteration is not reflected in the pollen diagram. The burning would have reduced the extent of rainforest on well-drained sites allowing secondary *Eucalyptus* spp. to expand, and would have increased the extent of wet scrub and heath on the poorly-drained flats and gently inclined slopes of the valley bottom.

DATING THE KING VALLEY GLACIAL SEQUENCE

This section outlines the chronology and age estimates of the glacial events in the King Valley. The tentative ages of the glacial formations recognised are summarised in Figure 40. This chronology is based on estimates of ages using several methods including ^{14}C dating, amino acid analyses, palaeomagnetic measurements and weathering data. Most of the estimates are minimum ages and should be regarded as first approximations.

The maximum age of the Long Marsh Formation has been determined by a radiocarbon date of 12 250 ± 90 yr B.P. (SUA 2415) from a large *Eucalyptus* branch beneath alluvial gravel. The date suggests that the upper part of the King Valley supported a *Eucalyptus* forest with relatively large trees by this time. The presence of large trees suggests that the valley was largely ice free by 12 000 yr B.P.

The age of the Dante Formation has been determined by ^{14}C dating that suggests it post dates 18 800 yr B.P. Although this date has been given the status of *the* date of the maximum of the Last or Margaret Glaciation, the section is not without its problems. The existence of unweathered, Last Glaciation tills (the Chamouni Formation, >48 000 yr B.P.) outside the limits of glacigenic sediments of the Dante Formation demonstrated that the Dante Formation does not necessarily record the maximum extent of ice during the Last Glaciation.

The 48 700^{+2900}_{-2100}yr B.P. (SUA 2599) date obtained on drifted wood that forms part of the Chamouni Formation outwash gravel suggests the age of the formation is beyond ^{14}C dating and therefore more likely to have been deposited by an ice advance that occurred more than 25 000 yr before the Dante advance. The Chamouni Formation is interpreted as being considerably younger than the Henty age formations of the King Valley because its deposits lie geographically closer to the sources of ice and the deposits are considerably less altered by chemical weathering. At present the dating of the Chamouni advance is equivocal but the field evidence suggests that it may represent an early Last Glaciation ice advance.

The age of the Smelter Formation has been determined by an amino acid analysis on wood from the base of the deposit that suggests a minimum age equivalent to oxygen isotope stage 5 (B. J. Pillans, pers. comm., 1988). Although the Smelter Formation cannot be directly related to the deposits of the King Glacier, they are regarded as being of last interglacial age and are correlated with the Pieman Interglaciation (Colhoun, 1980) of the northern part of the West Coast Range.

The age estimate of the Bull Rivulet Formation is dependent on its close geographic relationship, and inferred temporal relationship, to the David Formation. It is considered to be of Henty Glacial age though there is little exposure from which to deduce the characteristics of the deposits which may prove to be considerably younger.

There is no reliable absolute date for the David Formation, in the lower King Valley. The two ^{14}C dates of 32 800

$^{+400}_{-700}$ yr B.P. (SUA 2392) and 39 300$^{+800}_{-700}$ yr B.P. (SUA 2393) from the Crotty Plain are probably contaminated by modern carbon which suggests the samples are beyond the range of radiocarbon dating. The age of the sediments can be assessed by comparing the weathering rinds with those in the dated Dante Formation. The mean weathering rind thickness for the David Formation is 11.2 mm, which is about seven times that of the Dante Formation (1.5 mm). Assuming a linear weathering rate and given the age of the Dante Formation (18 800 yr) the weathering rind thicknesses suggest that the David Formation is at least 130 000 years old.

The age of the Cableway Formation can be estimated by assessing the period of time represented by the Nelson Formation that lies between the Cableway and David Formations. The period of time represented by the 44 m of Nelson Formation lake sediments have been estimated as representing about 4400 years assuming similar sedimentation rates to Holocene proglacial lakes in New Zealand. Because the top of the lake sediments are eroded this estimate is probably conservative. Taking the age of the David Formation as 130 000 yr, the Cableway Formation is at least 4000 years older than the David Formation. Although this estimate cannot be regarded as precise, it suggests that the deposits are certainly not younger than the last interglacial in age but are not likely to be extremely old.

The age of the Moore Formation can be estimated by comparing the thickness of weathering rinds with those of the Cableway Formation. This direct comparison is only possible because the contact between the Cableway and Moore formations can be observed in two places. At these sites the Cableway Formation has a mean weathering rind thickness of 5.2 mm and the Moore Formation has a mean value of 15.8 mm. The difference suggests that the Moore Formation is at least three times as old as the Cableway Formation. This gives a minimum estimated age of about 400 000 years. The magnitude of the estimated unconformity between the Cableway and Moore formations suggests they were deposited during different glaciations.

The Pyramid Formation was deposited during the same glaciation as the Moore Formation, and therefore is regarded as being of approximately the same age.

The Huxley Formation is separated from the Pyramid and Moore formations by the Baxter Interstadial deposit. The period of time represented by the Baxter Interstadial deposit is unknown. The time gap between the deposition of the Huxley and Pyramid formations cannot be estimated by weathering because the sediments are largely siliceous and contain no dolerite. The Baxter Interstadial sediments have a normal detrital remanent magnetisation which indicates that it belongs in the Brunhes Chron which is in accordance with age estimates for the Moore Formation. This assessment of age is supported by the amino acid date of wood from the Baxter Formation which suggested a minimum age equivalent to oxygen isotope stage 10 (B. J. Pillans, pers. comm., 1987).

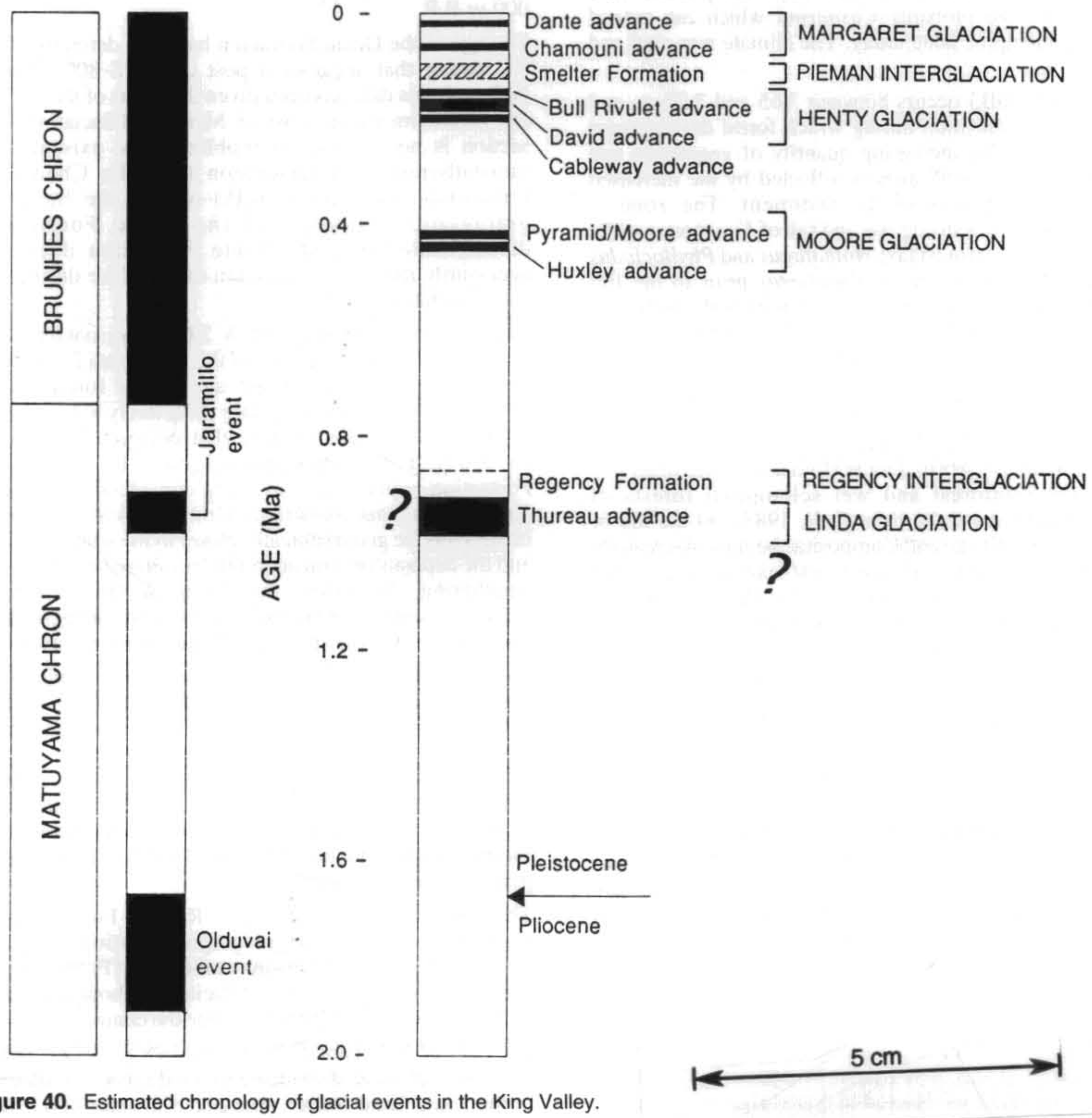


Figure 40. Estimated chronology of glacial events in the King Valley.

The age of the Thureau Formation is more difficult to estimate. It is known to have a reversed detrital remanent magnetisation and therefore is probably older than 730 000 yr. Because the sediments at the type section of the Thureau Formation appear to cross a palaeomagnetic boundary it may be that they contain part of the Jaramillo or Olduvai events in the Brunhes Chron, which occur at 890 000 and 1 620 000 yr B.P. (Bowen, 1978).

GLACIATIONS AND INTERGLACIATIONS IN TASMANIA

Developing a time-stratigraphic framework

Development of a time-stratigraphic framework for Quaternary glacial deposits in Tasmania is based largely on the experience of the New Zealand Geological Survey (Suggate, 1965*a*, 1965*b*, 1985*b*). Time-stratigraphic subdivision of the Quaternary is into stages and substages that are based on climatic criteria which offers a means of worldwide correlation (Suggate, 1965*a*). The critical points of climatic change are the commencement of cooling leading to glaciation and the commencement of warming leading to interglaciations. Because the evidence of terrestrial deposits is discontinuous and fragmentary the critical points of climatic change are difficult to find and identify. Consequently, in Tasmania stage definitions depend on age estimates derived from radiocarbon dating, palaeomagnetism, amino acid analyses and the degree of weathering of deposits.

In this discussion glaciation is used in both a climatic and in a stratigraphic sense. In the climatic sense, a glaciation is an event of cooling that leads to the formation of glaciers. In a stratigraphic sense, it is necessary to distinguish the event from the deposits that formed as a consequence of the event (Suggate, 1965*b*). A glaciation is a number of ice advances that occur in close temporal proximity to each other. A glacial climatic stage represents a period of cold climate during which one or more phases of ice advance occurs. Interglacials are non glacial periods in which the length and/or degree of climatic amelioration was sufficient to permit the development of forest indicative of climate at least as warm as was attained during the post-glacial climatic optimum (West, 1961). In the King Valley interglacials are recognised as biostratigraphic units defined by the presence of pollen or plant macrofossils that record the development of temperate rainforest.

Stadials are cooler periods within glaciations during which glaciers are more extensive. Stadials are separated by interstadials which are periods of climatic amelioration within glacial events (Bowen, 1978). Interstadial climates are generally considered to have been cooler than the Holocene optimum. During interstadials the temperature probably rose to about halfway between the minimum temperature of the Last Glaciation and the maximum warmth of the post glacial climatic optimum (West, 1961; Suggate, 1965*b*). In the King Valley, stadials are recognised by glacier advances and defined as morpho-stratigraphic and/or lithostratigraphic units. Interstadials are recognised as biostratigraphic and/or litho-stratigraphic units on the basis of having heath and shrub vegetation assemblages and/or accumulations of non-glacial sediments.

The pre-glacial environment

Recognition of the preglacial environment is based on the absence of evidence for glaciation. It can be considered a stable period of subtropical-temperate climate with a slow cooling trend from the middle to late Tertiary. One of the difficulties in discussing the pre-glacial environment is that the antiquity of glaciation in Tasmania is not known – it could be early Pleistocene or Late Tertiary. Diamictons recorded at Lemonthyme Creek (Paterson *et al.*, 1967) have been tentatively interpreted as tillites of Tertiary age by Colhoun (1975). These diamictons are overlain by organic silts and clays that contain a Tertiary temperate rainforest flora.

Tertiary pollen and macrofossils of tropical and subtropical plant taxa known in Miocene and Oligocene sediments (Gill, 1962) show evidence for higher temperatures than present. A cooling trend appears to occur in the Late Tertiary (Hill and Macphail, 1983). The plant fossil composition at Pioneer in north-eastern Tasmania describes a complex temperate rainforest dominated by *Nothofagus*. From a count of 618 pollen grains Hill and Macphail (1983) found that 47% were *N. emarcidus/heterus*, an extinct species of the *brassii* group. The Idaho palaeosol also contains forest taxa that are now extinct, including 13% *N. brassii* type pollen (Kiernan, 1980). At Regatta Point, about 20 km west of the King Valley, an eroded block of organic silt that is probably no older than Pliocene in age, contains pollen of temperate rainforest taxa some of which are now extinct in Tasmania (Hill and Macphail, 1985).

The only known preglacial sediments in the King Valley are the fluvial sediments of the Idaho Formation which are preserved as an inlier in Thureau Formation glacigenic sediments in the Linda Valley. In Tasmania, the presence or absence of extinct temperate rainforest species is often referred to as the difference between the Tertiary and the Pleistocene, and many floras such as that preserved in the Idaho Formation are claimed to have 'Tertiary affinity' (e.g. Kiernan, 1980; Colhoun, 1985*a*). The reality is that the Tertiary, and particularly the later part of the Tertiary, is characterised by a lack of knowledge about the composition or temporal range of the flora. At present it is not possible to estimate the age of deposits from floristic characteristics. The range of many of the species that are often claimed to have 'Tertiary affinity' is not known. The identification of *Quintinia psilatispora* and *Gothanipollis perplexus* in Pleistocene sediments in the King Valley has considerably extended their temporal range.

Little is known about the late Tertiary preglacial vegetation other than that many species became extinct during the late Tertiary and early Pleistocene. To what extent the extinction of numerous forest taxa recorded in late Tertiary sediments is related to cooling temperatures and/or glaciation remains unknown.

Lemonthyme Glacial Stage

Evidence for the Lemonthyme Glaciation consists of drill cores through sediments at Lemonthyme Creek in northern Tasmania. The cores show sequences of laminated muds overlying and interbedded with tillite (Paterson, 1965), and the silts are known to contain pollen that indicates cool temperate rainforest dominated by *Nothofagus* (Paterson *et al.*, 1967). Over half of the *Nothofagus* pollen grains are species that are now extinct, suggesting that the flora has closer affinity to those described from Pioneer (Hill and Macphail, 1983) and Idaho (Kiernan, 1980) than to any Quaternary interglacial flora.

Because the Lemonthyme Glaciation is known only from subsurface information at Lemonthyme Creek, neither the extent of ice, nor the environment at the time of deposition are known. No glacial deposits of this antiquity are known in the King Valley or elsewhere in Tasmania. Their age, origin and full significance remains to be evaluated.

Linda Glacial Stage

The term Linda Glacial Stage was first used by Kiernan (1983*a*), although the stage was not defined and the term 'stage' was used interchangeably with 'glaciation'. The

Linda Glacial Stage is here defined as including all deposits that formed after the beginning of the cooling that led to the Linda Glaciation and before the warming that led to the Regency Interglaciation. The type formation for the Linda Glaciation is the Thureau Formation in the King Valley. Deposits of the Linda Glaciation are clearly the most extensive glacial deposits in western Tasmania. This differs from the implications of the oxygen isotope records from the southern ocean, which suggest that maximum ice volumes were attained during oxygen isotope stage 6 (≈140 000 yr B.P.). The Linda deposits consist of a wide range of sediments dominated by flow tills, laminated lake sediments and ice-rafted tills. The lack of melt-out till and the relative abundance of outwash gravel and lodgement till suggests that the glaciers were of a temperate maritime type and the depositional environments were characterised by large fluxes of meltwater. The altitude attained by erratics on the hillsides suggests that the King Glacier was at least 400 m thick in the middle part of the valley near the Thureau Hills.

Although there is some suggestion of multiple ice advances in the Thureau Formation, evidence to substantiate this and suggestions that deposits of the Linda Glaciation contain deposits of several glaciations (Kiernan, 1983*b*; Gibson *et al.*, 1987) has not been forthcoming.

No glacial floras of Linda Glaciation age are known in Tasmania. Deposits of the Regency Interglacial conformably overlie the Thureau Formation and suggest that the local environment immediately following deglaciation was dominated by aquatic and subalpine taxa. The Regency Interglacial deposits are the only ones known to occur between the Linda Glaciation and subsequent middle Pleistocene ice advances.

Regency Interglacial Stage

The Regency Interglacial Stage is defined as including all deposits that formed after the beginning of the warming that led to the Regency Interglaciation and before the cooling that led to the Moore Glaciation. The type formation of the Regency Interglaciation is the Regency Formation. The Regency Interglacial consists of 0.9 m of drifted wood, leaves and organic silty sand that conformably overlie Thureau Formation outwash gravels. Pollen analysis of these sediments indicates the development of temperate rainforest dominated by *Nothofagus cunninghamii*, *Phyllocladus aspleniifolius* and *Lagarostrobos franklinii* (fig 16). The abundance of tree ferns *Cyathea* and *Dicksonia antarctica* (24%) in the upper part of the diagram suggests that the climate was at least as warm as the post-glacial climatic optimum, when the value for tree-ferns in the King Valley was considerably lower (fig. 18).

The Regency Formation pollen diagram can be compared to zones 5 to 3 of the Langdon Interglacial (Colhoun *et al.*, 1989) which can be used as a model of the structural characteristics of interglacial vegetation change and the successional establishment of temperate rainforest in western Tasmania. These zones are dominated by *Casuarina–Eucalyptus* (zone 5), *Phyllocladus–Casuarina* (zone 4), and *Phyllocladus–Casuarina–Bauera–Nothofagus* (zone 3). No such succession occurs in the Regency Interglacial. Prominent in the early stages of development of rainforest at Langdon River is the increasing abundance of *Phyllocladus* and near absence of *Nothofagus* as *Eucalyptus* and herbs decline. *Phyllocladus* continues to dominate as the amount of *Nothofagus* increases and *Casuarina* and *Eucalyptus* maintain stable values. *Phyllocladus* pollen in the Regency Interglacial maintains consistent values throughout the profile and never achieves the dominance that it does at Langdon River.

The greater abundance of *Cyathea* in the Regency profile compared to *Dicksonia* appears to have significance because in Holocene pollen diagrams (Macphail, 1979) *Dicksonia* is the dominant tree-fern. Towards the top of the Regency profile the appearance of the tree ferns and abundance of *Lagarostrobos franklinii* suggests the rainforest canopy was more open though the reason for it becoming open cannot be determined from the pollen analysis.

The Regency Interglacial, the oldest known interglacial flora from western Tasmania, can be contrasted with what are regarded to be Oligocene, Pliocene, and early Pleistocene palynofloras. An interesting comparison is that the amount of extinct *Nothofagus brassii*-type pollen found at several sites appears to decrease in the younger deposits (table 7).

The abundance of extinct species in organic silts overlying the Lemonthyme tillite, and the lack of extinct species in the Regency Interglacial except for traces of two extinct rainforest taxa *Gothanipollis perplexus* and *Quintinia psilatispora*, suggests that the Regency Interglacial and the interglacial temperate palynoflora at Lemonthyme are widely separated in time (table 7). If the decreasing amount of extinct taxa recorded in rainforest environments in Tasmania is a pattern that is due to age, then the age of the pollen sequences are as they are ranked in Table 7. However, as already noted, both the causes and the time of extinction of the numerous rainforest taxa have yet to be established. What is apparent is that the change is almost certainly due to a gradual extinction of species through time.

Table 7. EXTINCT *NOTHOFAGUS* SPECIES IN TASMANIAN RAINFOREST PALYNOFLORAS

Site	*% brassii-type pollen*
Pioneer[1]	47
Lemonthyme[2]	32
Gormanston[3]	13
Regatta Point[4]	trace
Regency	0

1. Hill and Macphail (1983)
2. Paterson et al. (1967)
3. Kiernan (1980)
4. Hill and Macphail (1985)

Moore Glacial Stage

The Moore Glacial Stage is defined as including all deposits that formed after the beginning of the cooling that led to the Moore Glaciation and before the warming that led to the Langdon Interglacial which was defined by Colhoun *et al.* (1989). The type formation of the Moore Glaciation is the Moore Formation. Evidence for the Moore Glaciation is limited to two sites which show major weathering unconformities between the Cableway and Moore formations. Because sediments of the Moore Glaciation transported by the King Glacier have no depositional forms and crop out only beneath younger deposits, it is not possible to reconstruct accurately the extent of ice during the Moore ice advance. However, the large particle sizes of the sediments suggest the ice limit was close to Baxter Rivulet and that its extent would have been similar to that of the Cableway advance (fig. 2).

The Moore Glaciation consists of two major ice advances. The outwash gravels of the ice advances are separated by organic-rich fluviatile sediments of the Baxter Interstadial. Pollen analysis of these sediments suggest there was a climatic amelioration between the advances. It is this inferred warming that is the basis for defining the Baxter Interstadial. The Baxter Interstadial records a wet heath (pollen zone BR 1) and a heath-herbland mosaic

(BR 2, fig. 17). The pollen assemblage is quite different from the palynofloras of other organic-rich deposits in its very high values of *Casuarina* and very low values of *Eucalyptus*. The same contrast has been noted between an interglacial pollen/vegetation succession at Langdon River which has a sequence of *Casuarina monilifera*, *Phyllocladus aspleniifolius* and *N. cunninghamii* (Colhoun *et al.*, 1988) and the Holocene succession of *Eucalyptus*, *Phyllocladus*, *Nothofagus* and *Eucryphia* (Macphail, 1979; Colhoun and van de Geer, 1986). This pollen analysis indicates a cool non-forested interstadial type climate, an interpretation that the overlying glacigenic deposits support. However, it is possible, though not determinable from the section, that the two pollen zones BR2 and BR1 may represent the final part of an interglacial sequence as climate was deteriorating.

Accepting that the interstadial interpretation is correct, then it is likely that mean annual temperature at the time of formation of the Baxter Interstadial sediments would have been considerably lower than present. Since the composition of the palynoflora suggests a treeline close to the altitude of Baxter Rivulet (≈180 m), a temperature depression of 4–5°C is implied because present treeline occurs at about 1000 m.

Langdon Interglacial Stage

The Langdon Interglacial is known from pollen and plant macrofossils that occur outside the limits of the Last Glaciation and overlie glacial deposits at Langdon River, about 12 km north-west of the King Valley. The Langdon Interglacial records the successional development of temperate rainforest as indicated above (Colhoun *et al.*, 1989). Amino acid dating of wood from the Langdon organic deposit indicates that it was deposited during oxygen isotope stage 7 (B. J. Pillans, pers. comm., 1988). No equivalent deposits are known from the King Valley.

Henty Glacial Stage

The term Henty Glacial Stage was first used by Colhoun (1979) for till exposed at Henty Bridge that was considered to predate the last interglacial. Kiernan (1983) also used the term, although it was not defined and was used interchangeably with the term 'glaciation'. The Henty Glacial Stage is defined here as including all deposits that formed after the beginning of the cooling that led to the Henty Glaciation and before the warming that led to the Pieman Interglaciation. The type section for the Henty Glaciation occurs at Henty Bridge, west of the West Coast Range. Deposits of the Henty Glaciation are the most widespread glacial deposits in the King Valley where it is represented by three advances the latest two of which (the Bull Rivulet and David advances) may be separated by a very short period of time. Separation of the Cableway and David advances is based on the presence of over 40 m of non-glacial lake sediments between them.

At the maximum extent of the Henty Glaciation the King Glacier was over 300 m thick and almost as extensive as during the Thureau advance. Geological processes at this time, though dominated by glacial action, also included widespread periglacial action which led to the deposition of thick angular screes that are interbedded with and overlie Henty Glaciation sediments. On the eastern slopes of Mt Owen, parts of these screes became solifluction lobes which extend down onto the floor of the valley and across the David Formation end moraine. The widespread nature of periglacial deposits suggest that during and after ice retreat from the extensive David and Cableway advances, the climate remained cold for some time before the slopes became vegetated.

The Nelson Interstadial is defined by the absence of diamictons in the 44 m of lake sediments rather than the presence of an interstadial flora. The lack of dropstones and glacigenic deposits in these sediments suggests the sediments accumulated at the bottom of a non ice-contact lake and that at least the middle part of the King Valley was free of ice at this time. On this basis, the sediments must be regarded as of interstadial character because they represent a comparatively long-lived non glacial period in a glacial event, even though the climate may have been cold.

Pieman Interglacial Stage

The Pieman Interglacial Stage is defined as including all deposits that formed after the beginning of the warming that led to the Pieman Interglaciation and before the cooling that led to the Margaret Glaciation. Organic sediments at Smelter Creek [CP865289] record a temperate rainforest flora which appears to have been deposited during the last interglacial. Amino acid dating of wood from the Smelter Creek organic deposit indicates that it was deposited during oxygen isotope stage 5 (B. J. Pillans, pers. comm.). On this basis it appears that the deposits are a correlate of the interglacial deposits at Pieman dam site described by Colhoun (1980).

Margaret Glacial Stage

The term Margaret Glacial Stage was first used by Colhoun (1979) synonymously with the Last Glacial Stage and based on Lewis' identification of the Margaret Glaciation (Lewis, 1945). The term Margaret Glacial Stage was also used by Kiernan (1983) although it was not defined and the term 'stage' was used interchangeably with the term 'glaciation'. The Margaret Glacial Stage is defined here as including all deposits that formed after the beginning of the cooling that led to the Margaret Glaciation and before the warming that led to the post glacial period. The type formation of the Margaret Glaciation is the Dante Formation Gibson *et al.*, 1987). In the King Valley, deposits of the Margaret Glaciation are limited to the upper part of the valley near Dante Rivulet. Prior to this study the Last Glaciation was thought to be relatively simple, with one advance that occurred about 18 000 years ago (Kiernan, 1983*b*; Colhoun, 1985*b*). However, recognition of the Chamouni Formation as being of Margaret Glaciation age and predating the Dante Formation suggests the stratigraphy of the Margaret Glaciation is more complex.

The main issue is that tills of the Chamouni Formation of Margaret Glaciation age lie up to 4 km outside the limits of the Dante advance. This implies that the climate was cool enough for glaciers to form during the Margaret Glaciation prior to 21 000 yr B.P. Although the evidence for an earlier ice advance during the Margaret Glaciation is compelling for the King Valley, no similar advance has yet been recorded elsewhere in the West Coast Range.

Pollen from the silts that conformably rest beneath the Chamouni Formation record an alpine-subalpine assemblage that is consistent with the onset of the Chamouni advance (table 6).

Pollen analysis of the alpine humus palaeosol that underlies the Dante Formation outwash gravel (fig. 37) suggests that a cold climate herbfield-bog mosaic existed prior to the Dante advance (Gibson *et al.*, 1987). A similar palynoflora is recorded in the lower King Valley which was not glaciated during this time (fig. 39). The late glacial part of this flora, preserved in a filled sinkhole, is dominated by subalpine and alpine trees and shrubs.

Ice was absent from the floor of the valley during most of the Margaret Glaciation. The major geomorphological changes were caused by outwash streams and periglacial processes acting on deforested slopes. The degree to which periglacial processes modified the landscape at this time is difficult to assess because the deposits cannot be distinguished from older periglacial slope mantles.

However, the periglacial deposits in the King Valley are believed to have formed mainly during and subsequent to the Henty Glaciation. One of the major geomorphic effects of the Margaret Glaciation was the entrenchment of the King River into thick middle Pleistocene glacial deposits that cover much of the valley floor.

The transition from the Margaret Glaciation to the Holocene appears to have been abrupt (fig. 39) and occurred around 13 000 yr B.P. The date for the filled sinkhole (13 010 ± 130 yr B.P. SUA 2723) is confirmed by the presence of tree trunks in Holocene gravels in the upper King Valley dated at 12 250 ± 90 yr B.P. (SUA 2415), suggesting that large trees were in the valley at this time and that ice was absent. These minimum dates for the deglaciation of the King Valley are slightly older than most other sites for the West Coast range area, e.g. Poets Hill (11 420 ± 700 yr B.P. Gak 1826, Davies, 1974) and Tullabardine Dam (11 660 ± 150 yr B.P. SUA 1044, Colhoun and van de Geer, 1986).

REGIONAL CORRELATION

Correlation of sediments and rocks aims to demonstrate correspondence in character and/or stratigraphic position (Hedberg, 1976). This discussion concerns temporal correlation.

In order to demonstrate temporal correspondence the age of the correlated sediment bodies must be known. When correlation precedes dating it is difficult to argue about contemporaneity let alone about age differences (Vita-Finzi, 1973). Having recognised this, and considering the lack of securely dated sediments beyond the range of ^{14}C dating in Tasmania, any attempt at correlation of glacial events requires several assumptions. The most important of these is that the major glacial events were broadly synchronous.

To attempt a correlation of glacial events within Tasmania requires an assessment and comparison of the chronologies of other studies. Such a comparison is beset by several problems that include a lack of relative dating data, absence of absolute dates for deposits older than the late part of the Last Glaciation, and lack of a consistent approach to field mapping and stratigraphic classification.

A suggested correlation of the major glacial sequences in Tasmania is summarised in Table 8. The correlations have been made on the basis of radiocarbon dating for the late Last Glaciation, analysis of weathering rind data from Jurassic dolerite clasts, stratigraphic and geographic relationships and other dating methods for the older deposits.

Though there were several early studies that attempted to describe the glaciation of the West Coast Range area (see Banks *et al.*, 1987), and multiple glaciation was suggested by Lewis (1945), the more recent studies of the area (Ahmad *et al.*, 1959; Derbyshire, 1963; Read, 1963) suggested that most of the data previously observed could be explained as the product of a single glacial cycle. Since none of these studies was based on detailed stratigraphic analyses and did not have the availability of modern dating methods they do not contain data relevant to the correlation of glacial events in the King Valley.

The nomenclature of the glaciations has a complex history. The term 'Margaret' for the Last Glaciation has been preserved from the early work of Lewis (1945). The term 'Henty' comes from the recognition by Banks *et al.* (1977) that the till at Henty bridge predated the Last Glaciation. The till compares with that described by Lewis at Yolande bridge where it cannot be dated. The terms 'Dante' and 'Comstock' were used by Kiernan (1980) for Last Glaciation and pre Last Glaciation drift on the eastern side of the West Coast Range. Both Dante and Comstock have not been used in this report because it appears that they are the equivalents of Margaret and Henty. The term Linda was adopted because the deposits in the Linda Valley were originally described by Lewis as being of the same age as those at Malanna which have since been shown to be non-glacial (Banks and Ahmad, 1959). Since the Linda Valley was evidence for the oldest glaciation in Tasmania the name was selected to recognise its priority.

Studies with a greater emphasis on the stratigraphy of the glacial sediments began in the 1970s with Bowden's (1974) and Sansom's (1978) reconnaissance studies of glaciation in parts of western Tasmania. Although they found evidence for multiple glaciation, they do not have a consistent approach to stratigraphic classification or to field mapping. A later (1980) thesis by Kiernan, and two papers arising from it (Kiernan, 1983*a*, 1983*b*), were early studies that addressed questions of the stratigraphy and antiquity of Tasmania's glacial sediments. The area he studied, the central part of the West Coast Range, extends into the King Valley.

Kiernan's work provided data that confirmed the occurrence of three glacial stages, the Dante (equivalent to the Margaret), Comstock (equivalent to the Henty), and Linda glaciations. The importance of these stages is confirmed by the present study. Kiernan differentiated the stages on the basis of differences in weathering rind thickness on Jurassic dolerite clasts and radiocarbon dating of the late Last Glaciation. Large differences in weathering rind thicknesses were interpreted as representing interglacial warming.

The Dante 'Glaciation' (Kiernan, 1983*a*), is clearly the equivalent of the last or Margaret Glaciation (table 8). The Dante Formation is regarded as a late Last Glaciation advance of the King Glacier and has formation rather than glaciation status. The Chamouni Formation represents an early Last Glaciation advance of the King Glacier up to 4 km beyond the limits of the Dante sediments, and does not correlate with any events previously identified.

The Comstock Glaciation defined by Kiernan as a probable correlate of the Henty Glaciation east of the West Coast Range includes deposits defined in this study as Chamouni, Bull Rivulet, David and Cableway formations. Although Kiernan (1980) subdivided deposits of his Comstock glaciation into four parts, the divisions were based on altitude and geographic position rather than on stratigraphic position. The divisions therefore cannot be related to the glacial advances identified in this study. The name Comstock should be dropped in favour of Henty because all known evidence indicates that the deposits belong to the same period of glaciation.

The term Linda Glaciation became used informally for those deposits at the head of the Linda Valley, which Lewis (1945) regarded as correlates of the Malanna Glaciation, after Banks and Ahmad (1959) demonstrated that the deposits at Malanna were non-glacial. Kiernan (1983*b*) studied the weathering rinds on these Linda deposits and showed that they were very much more weathered than his Comstock deposits. He referred to them as deposits of the Linda Glacial Stage that preceded the Comstock Glaciation. Subsequently it has been shown that the Linda Glaciation lake sediments have a reversed detrital remanent magnetisation. Essentially the same criteria have been used for its identification in this study. In addition, the recognition of the organic sediments of the Regency Formation demonstrate that the Linda Glaciation was followed by interglacial warming and the development of temperate rainforest. The subdivisions that Kiernan (1980) made within the Linda Glaciation deposits (Linda 1 and Linda 2) cannot at present be demonstrated to belong to different glacial advances.

Augustinus (1982) and Augustinus and Colhoun (1986) report a study of the glacial stratigraphy of the Pieman Valley in the northern part of the West Coast Range about

Table 8. CORRELATION OF QUATERNARY GLACIAL SEQUENCES IN TASMANIA

Glaciations	*Interglacials*	*King Valley (this study)*	*West Coast Range (Colhoun 1979, 1985a)*	*Central West Coast Range (Kiernan, 1980)*	*Pieman Valley (Augustinus, 1982)*	*Ben Lomond (Caine, 1983)*	*Lake St Clair, central plateau (Kiernan, 1985)*	*Mersey Valley (Hannan and Colhoun 1987)*
Margaret		Dante	Margaret	Dante	—	Cirque	Cynthia Bay	Rowallan
		Chamouni					—	—
	Pieman	—	—	—	—	—	—	—
Henty		Bull Rivulet Blackwood King	Henty	Comstock	? Boco 2 ? Boco 1	Plateau	Beehive Powers Creek Clarence	Arm
Governor		Governor / Fish			? Boco 1	—	—	—
		Traveller	—			—	—	—
	Regency	Regency		—	—	—	—	—
Linda		Thureau	Linda	Linda	? Bulgobac ? Que	—	Stonehaven	Croesus
	Lemonthyme interglacial	—	—	—	—	—	—	—
Lemonthyme		—	—	—	—	—	—	—

35 km north-west of the King Valley. Glacial deposits were differentiated as 'drift sheets' that 'probably represented first order glacial events'. Although the meaning of a first order glacial event is unclear, it is used in a manner that suggests the deposits were divided into glaciations that were differentiated primarily by their weathering characteristics. Augustinus (1982) identifies two glacial stages, but Augustinus and Colhoun (1986) identified four glacial stages with the same evidence. The four glacial stages are Boco 2, Boco 1, Bulgobac and Que glaciations. It is intriguing to note that all glaciations known from the Pieman Valley consist of single ice advances whereas up to three distinct ice advances in a single glaciation have been identified in the King Valley. Because the interpretation of the Pieman glacial sequence is uncertain, correlations with the King Valley are suffixed with question marks (table 8).

Caine's (1983) study of the glaciation of the Ben Lomond massif suggested that there were deposits that related to two separate glaciations which he named the Cirque and Plateau. These are correlated with other glacial sequences in Tasmania on the basis of a count back and weathering characteristics of Jurassic dolerite (table 8).

Kiernan's (1985) study of the glacial deposits of the Central Plateau is the most extensive study of glaciation in Tasmania. The size and lack of exposure in the study area required an approach that synthesised the mainly morphological evidence from several catchments. The essentially morphostratigraphic study used moraines and outwash surfaces as primary mapping units. Relative dating of the deposits was based primarily on weathering rinds on Jurassic dolerite, but also included a wide range of criteria related to post-depositional modification of landforms and sediments.

Kiernan recognised three weathering zones that are surficial rock units delimited from others by their weathering characteristics (Boyer and Pheasant, 1974). Weathering zone 1 includes deposits of the Stonehaven drift which Kiernan suggests is a correlate of the Linda Glaciation. Weathering zone 2 includes the Beehive, Powers Creek and Clarence sediments which are collectively known as the Butlers Gorge Complex and have not yet been correlated with the deposits of the West Coast region. Weathering zone 3 consists of the little-weathered Cynthia Bay sediments of Last Glaciation age.

Although Kiernan applies three weathering rate curves to estimate the ages of these deposits, the estimates differ widely. They suggest the comparison of weathering characteristics of widely dispersed sites is not sufficiently accurate to allow a true appreciation of the relative age of all deposits.

The weathering of Cynthia Bay sediments suggests they are a correlate of the Margaret Glaciation. On the basis of counting backwards and weathering rinds, the Butlers Gorge Complex may be a correlative of the Henty and/or Moore glaciations. Any more detailed correlation of individual advances is not possible without being able to obtain absolute ages of the sediments.

The advanced state of weathering of the Stonehaven sediments with respect to that of the Butlers Gorge Complex suggest that it is about the same age as Thureau Formation sediments in the King Valley (table 8). A major difficulty in correlation based on weathering indices occurs when chronologies become more refined and small differences in weathering are used to distinguish deposits of different ages. Small numerical differences in weathering cannot be accepted as a means of long distance correlation or for differentiating glacial advances. It is for this reason that the glacial stratigraphy of Tasmania has not become significantly more detailed or refined since the initial application of weathering as a relative dating method by Kiernan (1980, 1983*b*). This has occurred despite the subsequent intensive effort put into relative dating of deposits by weathering indices (Augustinus, 1982; Kiernan, 1985; Colhoun, 1985*b*; Augustinus and Colhoun, 1986).

Regional correlation based on weathering is bound to lead to errors and inconsistencies, yet it is the only tenable method available at present. Until the sediments are dated, it will not be possible to prove temporal correspondence of glacial events, and suggestions about correlations will remain speculative.

CORRELATION WITH GLACIAL EVENTS IN THE SOUTHERN HEMISPHERE MIDDLE LATITUDES

The record of Pleistocene glaciation of the King Valley is one of the more complete records of glaciation in the Southern Hemisphere, and is therefore of significance for hemispheric comparisons of glacial events. Western Tasmania, Chile, and Westland, New Zealand, are at similar latitudes, have similar west coast maritime climates and a similar topography, though the altitude of the West Coast Range of Tasmania is considerably lower than the Andes and the Southern Alps. Because the environments are similar, close similarities in responses to climatic change, at least during the Late Pleistocene, are to be expected. While it may be possible to assume that the major ice expansions in different parts of Tasmania were synchronous, assumptions about the contemporaneity of glacial events between Tasmania and similar regions in New Zealand and southern South America achieve little when one of the purposes of international correlation is to determine if episodes of terrestrial ice expansion were contemporaneous.

Several early attempts at correlation of Tasmanian glacial sequences with those of South America are summarised by Table 9. Kiernan (1980, 1983*b*) suggested that the Dante Glaciation was a correlate of the Kumara-2_2 advance of the Otira glaciation defined by Suggate (1965*a*), and the Late Llanquihue Glaciation in South America (Mercer, 1976). In the later paper, Kiernan also suggested the Comstock glaciation, which he was now calling the Henty Glaciation, may be a correlate of the Kumara-2_1 advance of the Otira Glaciation and/or the Waimea or Waimaunga glaciations (table 9). Although in the 1980 study Kiernan did not attempt to correlate the Linda Glaciation, in 1983 he inferred a correlation between it and the Porika Glaciation (table 9), though he also suggested the Linda Glaciation may be a great deal older.

Colhoun (1985*a*) suggested that the maximum of the Margaret Glaciation, as dated at Dante Rivulet, correlates with the Kumara-2_2 advance in Westland, the late Llanquihue glaciation in South America, and the peak of stage 2 of the oxygen isotope record from Indian Ocean cores. He regarded this as evidence which demonstrates the synchroneity of the Last Glaciation Maximum in west coast climates of the Southern Hemisphere. Although Colhoun makes no further correlations with the New Zealand glacial sequence, he suggests correlations with the South American sequence on the basis of similar weathering rind thicknesses. Specifically, he suggests that Margaret Glaciation deposits, as well as correlating with the late Llanquihue deposits, may also correlate with Llanquihue I deposits described by Mercer (1976), and Henty Glaciation deposits may correlate with Santa María, Río Llico and Caracol drifts on the basis of similarity in the thickness of weathering rinds on volcanic clasts in South America and Jurassic dolerite in Tasmania.

Kiernan (1985) suggested correlations between Cynthia Bay sediments and the Kumara-2_2 advance in New Zealand and the maximum of the Llanquihue in South America. Assuming broadly similar hemispheric responses to climatic change, he correlated the Butlers Gorge Complex with the Waimea and or Waimaunga glaciations in Westland (table 9) and pre Llanquihue glacial deposits in South America, and Stonehaven sediments with the Porika glaciation in New Zealand which was believed at that time to be about 750 000 years old (Mildenhall and Suggate, 1981).

Since these correlations were suggested the status of the New Zealand glacial chronology has changed significantly and now both the Porika and Ross glaciations are thought to be of Pliocene age (Suggate, 1985*b*). The change in the New Zealand chronology forces a reassessment of the correlations that have been suggested. It also raises the question of whether correlations of the glacial stratigraphy of Tasmania should be adjusted to accommodate changes in its supposed international

Table 9. SUGGESTED CORRELATIONS OF TASMANIAN GLACIATIONS WITH THE GLACIAL SEQUENCE OF NEW ZEALAND

<table>
<tr><th colspan="2">Suggate (1965a)*</th><th>Kiernan (1980)</th><th>Kiernan (1983b)</th><th>Augustinus (1982)</th><th>Colhoun (1985b)</th><th>Kiernan (1985)</th></tr>
<tr><td rowspan="3">Otira Glaciation</td><td>Kumara 3 (Moana)</td><td>—</td><td></td><td>—</td><td>—</td><td></td></tr>
<tr><td>Kumara 2² (Larrikins)</td><td>Dante Glaciation</td><td>Dante Glaciation</td><td>—</td><td>Margaret Glaciation maximum</td><td>Cynthia Bay</td></tr>
<tr><td>Kumara 2¹ (Loopline)</td><td>—</td><td>?</td><td>—</td><td>—</td><td rowspan="3">Butlers Gorge Complex</td></tr>
<tr><td colspan="2">Waimea Glaciation</td><td>Comstock Glaciation</td><td rowspan="2">Henty Glaciation</td><td>Boco 2 Glaciation</td><td>—</td></tr>
<tr><td colspan="2">Waimaunga Glaciation</td><td>—</td><td>Boco 1 Glaciation</td><td>—</td></tr>
<tr><td colspan="2">Porika Glaciation</td><td>—</td><td>Linda</td><td>Bulgobac Glaciation</td><td>—</td><td>Stonehaven</td></tr>
<tr><td colspan="2">Ross Glaciation</td><td>—</td><td>—</td><td>Que Glaciation</td><td></td><td>—</td></tr>
</table>

* Revised nomenclature of Suggate (1985*a*) in parentheses.

correlates. To suggest that it should is ridiculous, although the question highlights the dangers of correlating without dating, and emphasises that the purpose of correlation is to demonstrate temporal correspondence. Correlations suggested in Table 9, other than that of the dated late Margaret Glaciation maximum, assume synchroneity in the responses to climatic change and development of land ice in the Southern Hemisphere.

Despite the problems of the correlations discussed above, several observations about the timing of glaciations in the Southern Hemisphere can be made:

Intrahemispheric correlations of glacial events that are beyond the range of radiocarbon dating are hampered because of different preservation of deposits and because the three areas experienced very different uplift rates. Different preservation of deposits is dependent on postdepositional erosion, which can result in absence of parts of the depositional record of glaciation. Uplift rates of mountains can have an impact on the record of glaciation because glaciation in the Southern Hemisphere middle latitudes is dependent on the presence of mountain barriers that form the accumulation areas for glaciers. The formation and uplift of mountain barriers can therefore be a trigger to glaciation that is partly independent of climatic change. In New Zealand, lack of evidence for glaciation between 2.1 and 0.35 Ma has been attributed to the recency of uplift of the Southern Alps (Bowen, 1978) and to the role of uplift in causing erosion of evidence of glaciation (Suggate, 1985*b*). Consequently, glacial events in one area may have no counterparts in the other (Mercer, 1983). For these reasons, Tasmania, which was tectonically stable during the Quaternary, may prove to be a useful reference point in trans-Pacific correlations.

The pattern of glaciation in South America and Tasmania was broadly similar, with the greatest ice advances occurring in the Early Pleistocene and a succession of ice advances of decreasing magnitude in the Middle and Late Pleistocene. Although there is some suggestion that the greatest Patagonian glaciation (Mercer, 1983) and the Linda Glaciation may have occurred at similar times, lack of reliable dating of the Tasmanian sequence precludes correlation. It is intriguing to note that the greatest glacial events in Tasmania and South America that occurred in the Early Pleistocene are apparently not recorded in New Zealand, where there is no record of glaciation between 2.1 Ma and 350 000 yr B.P. (Suggate, 1985*b*).

The Regency interglacial deposit is the oldest known interglacial in Tasmania, and records the presence of temperate rainforest that suggests the climate was at least as warm if not warmer than the Holocene climatic optimum. No correlative deposits are known from South America or New Zealand.

The importance of the Moore glacial deposits to the glacial stratigraphy of Tasmania is that they record a multi-phase glaciation that appears to have occurred between 400 000 and 730 000 yr B.P., and an interstadial pollen flora preserved between glacial deposits. Although the age of the Moore Glaciation is not well constrained, there are no known comparable deposits in southern South America or New Zealand.

The minimum age estimate of 130 000 yr B.P. for the Henty Glaciation suggests it is the penultimate glaciation that occurred during oxygen isotope stage 6. The two main phases of the Henty Glaciation, the Cableway and David advances, may represent the two cold spikes of the oxygen isotope record from the oceans and the Antarctic ice sheet (Shackleton and Opdyke, 1973; Jouzel *et al.*, 1987). In Westland, the Waimea Glaciation (Suggate, 1965, 1985*a*, 1985*b*) also appears to be the penultimate glaciation, though lack of reliable dating of both events precludes correlation at this time. Although Colhoun (1985) suggested that the Santa María, Río Llico, and Caracol drifts of Chile (Porter, 1981) are possible correlates of the Henty Glaciation, the Chilean drifts are likely to be of at least early Middle Pleistocene age (S. C. Porter, pers. comm., 1990). However, lack of age estimates on the Chilean drifts means that firm correlations with glacial deposits in New Zealand or Tasmania are not possible at present.

The pollen assemblage at Smelter Creek represents the development of cool temperate rainforest during the Pieman Interglaciation. Amino acid analysis of wood from the sediments suggests deposition during isotope stage 5, and the stratigraphic position suggest the deposit represents the last interglacial stage. On this basis the Pieman Interglaciation seems to be a correlate of the Oturi Interglaciation, now the Kaihinu Interglaciation (Suggate, 1985*a*), of northern Westland.

The stratigraphic position and the ^{14}C date of 48 700 $^{+2900}_{-2100}$yr B.P. (SUA 2599) from the Chamouni Formation suggest it is an early last glaciation ice advance that may have occurred during oxygen isotope stage 4. In Chile, Porter (1981) described the Llanquihue I drift as the outermost moraine arc of the last glaciation. Although the age of the Llanquihue I deposits has yet to be resolved, ^{14}C dates indicate that the deposits may have formed during an ice advance between 30 000 and 19 000 yr B.P. or at a time beyond the limits of conventional radiocarbon dating (Stuiver *et al.*, 1975; Porter, 1981). In northern Westland, Suggate (1985*a*) suggests that the Loopline Formation (Loopline 1 of Suggate, 1965), represents an early last glaciation ice advance. The Chamouni Formation of the King Valley may correlate with the Llanquihue I drift of Chile and the Loopline Formation of north Westland. Although the ages of all three events are not yet well constrained, our interpretation of the Chamouni Formation lends support to other evidence of an early last glaciation ice advance and to suggestions that glaciers were largest during the last glaciation at a time close to or beyond the limits of radiocarbon dating (van der Hammen, *et al.*, 1981; Mercer, 1983).

Radiocarbon dates and the stratigraphic position of the Dante Formation suggest deposition during the peak of oxygen isotope stage 2. On the basis of ^{14}C dates, correlations between late last glaciation events in Tasmania, New Zealand, and South America have been suggested by Colhoun (1985). The ^{14}C dates of 18 800 ± 550 yr B.P. (ANU 2533) and 19 000 ± 170 yr B.P. (SUA 2856) on the Dante Formation (fig. 10) suggest correlation with the Larrikins Formation of Suggate (1985*a*), (Loopline 2 Formation of Suggate, 1965), and the Llanquihue II drift of central Chile (Porter, 1981). Colhoun (1985) calculated the equilibrium line-altitude (ELA) for the glaciers of the Margaret Glaciation at 42°S using the method described by Porter (1975). The ELAs were calculated to be 835 m for the small ice cap on the Tyndall Plateau, and between 800 and 1000 m for the cirque glaciers of the West Coast Range. Calculation of the modern snowline between 1815 and 1830 m indicates a snowline depression of about 1000 m for the late Margaret Glaciation. Comparable calculations have been made by Porter (1975, 1981) for 44°S in New Zealand and at 41°S in Chile. In New Zealand Porter (1975) reconstructed the ELA of glaciers in the Tekapo area and estimated a snowline depression of 875 m for the maximum of the last glacial maximum. In Chile Porter (1981) estimated the Llanquihue ELA at 900 and 925 m and inferred a snowline depression of 1000 m from the modern snowline at 1900–2250 m. The calculations from the three Southern Hemisphere middle latitude areas are all similar and indicate a snowline depression for the last glaciation of about 1000 m.

Mercer (1983) pointed to a further correspondence in glacial behaviour in New Zealand and southern South America with ice advances between 14 500 and 14 000 yr

B.P. In southern South America, Mercer (1976) and Porter (1981) infer a readvance of ice between 15 000 and 13 000 yr B.P. from the sedimentary record of water level fluctuations of Lago Llanquihue. In north Westland, a small readvance occurred between 14 500 and 14 000 yr B.P. (Suggate, 1965; Suggate and Moar, 1970). In contrast, there is no geological or palynological evidence in the King Valley, or elsewhere in Tasmania, for a readvance after the last glacial maximum (≈18 000 yr B.P.). Radiocarbon dates from logs in postglacial gravel and bogs suggest rapid deglaciation from 13 000 yr B.P. and the absence of ice from the upper King Valley by 12 250 yr B.P. Pollen evidence from a bog at the confluence of the King and Governor rivers (fig. 1) records an abrupt change from alpine heathland, *Microstrobos* scrub and heath at 13 000 yr B.P. to *Eucalyptus* subalpine woodland, succeeded by *Phyllocladus–Nothofagus* rainforest after 12 500 yr B.P. (Colhoun *et al.*, unpublished). Although the rainforest degenerated into scrub rainforest after 11 000 yr B.P., there is no evidence of cooling after the last glacial maximum. While evidence for a readvance after the last glacial maximum may yet be found in Tasmania, it appears that the climatic threshold for the development and maintenance of glacier ice was crossed by 12 000 yr B.P.

Comparison of Quaternary glacial events in Tasmania, New Zealand and South America shows that responses to climatic change in the southern hemisphere middle latitudes were broadly synchronous during the last glaciation. The Dante, Larrikins and Llanquihue II ice advances also appear to correlate with the greatest expansions of the Northern Hemisphere ice sheets during the last glacial maximum. However, evidence from the Southern Hemisphere of a more extensive ice advance during the early last glaciation contrasts with evidence from the Northern Hemisphere where the ice sheets reached their maximum extent after 18 000 yr B.P. (Denton and Hughes, 1981). Comparison of pre-Late Pleistocene events are hampered by lack of reliable dates on the deposits. Until the dating of deposits in all three middle latitude Southern Hemisphere areas is improved, correlations of glacial events will remain relatively speculative.

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