Buried Inset-Valleys in the Eastern Yilgarn Craton, Western Australia: Geomorphology, Age, and Allogenic Control

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ABSTRACT

A contributory network of narrow, fluvially incised valleys filled with a distinctive sequence of Tertiary continental to shallow marine sediments is widespread throughout the eastern Yilgarn Craton in southwestern Australia. Previously referred to as "deep leads," or "paleochannels," these are herein termed "inset-valleys" to more accurately reflect their geomorphic expression and emphasize their subordinate and entrenched position with the bedrock surface of a much broader and older system of "primary valleys." Owing to minimal denudation and tectonic deformation during the Late Cenozoic, the inset-valleys are excellently preserved and can be readily traced in boreholes and open-cut mines throughout the eastern Yilgarn Craton and possibly also adjacent parts of southwestern Australia. With a combined catchment areal extent of at least 1.2×10^6 km^2, this makes the inset-valleys among the largest of paleovalley systems known globally. Six major inset-valley networks occur within the eastern Yilgarn Craton, each comprising up to seven or more orders of tributaries arranged in a subdendritic pattern controlled by the initial slope provided by the primary valleys and, to a lesser degree, by lithological and structural discontinuities in the weathered Precambrian bedrock. The inset-valleys typically have a broad U-shaped form, a width-to-depth ratio of 10–15, and a maximum width and depth of about 1.5 km and 120 m, respectively. Inset-valley incision occurred in the early Middle Eocene in response to lowered geomorphic baselevel and increased stream gradients affected by epeirogenic uplift of the Yilgarn Plateau. Promoting inset-valley incision was a widespread mantle of deeply weathered bedrock and less certainly a marked reduction in fluvial sediment supply induced by climate change at the Middle/Late Eocene boundary.

Introduction

A well-integrated contributory network of narrow, bedrock-bounded, fluvially incised valleys filled with a distinctive sequence of Tertiary continental to shallow marine sediments is widespread throughout the eastern Yilgarn Craton in southwestern Australia (Clarke 1993; Kern and Commander 1993; Johnson et al. 1999). Previously known as "deep leads" to pioneering gold miners (e.g., Blatchford 1900) and "paleochannels" to subsequent workers (e.g., Kern and Commander 1993), they are herein termed "inset-valleys" (new term; see "Nomenclature" section) to more accurately reflect their size and shape and to emphasize their entrenched and subordinate position within the bedrock surface of a preexisting network of much broader "primary valleys."

Forming significant resources of gold, uranium, and groundwater and the most comprehensive record of Tertiary sedimentation on the Yilgarn Craton, the inset-valley fills have been studied in considerable detail (Blatchford 1900; Balme and Churchill 1959; Jones 1990; Clarke 1993, 1994*a*, 1994*b*; Kern and Commander 1993; Johnson and McQueen 2001; Anand and Paine 2002; de Broekert 2002). The inset-valley forms, by contrast, remain poorly described and understood, even though the episode of fluvial incision leading to their development constitutes a major event in the region's geological history and reflects a marked change in one or more of the main external or "allogenic" controls on erosion-deposition, which are tectonics, climate, and eustasy (Schumm and Ethridge 1994; Shanley and McCabe 1994; Ethridge et al. 1998; Blum and Törnqvist 2000).

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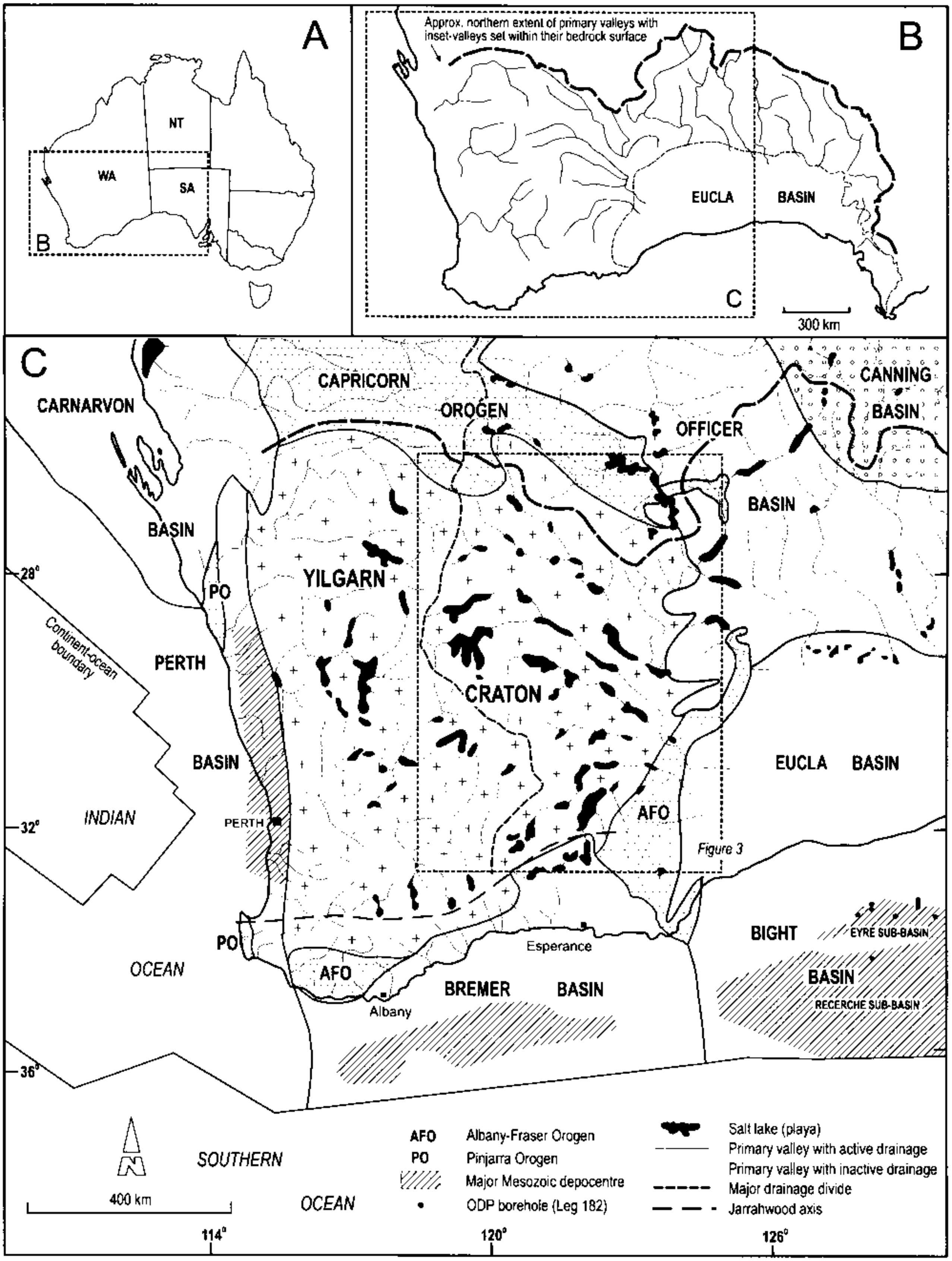

A
NT
WA
SA
B
Approx. northern extent of primary valleys with inset-valleys set within their bedrock surface
EUCLA
BASIN
C
300 km
C
CARNARVON
BASIN
CAPRICORN
OROGEN
OFFICER
BASIN
CANNING
BASIN
YILGARN
CRATON
PO
PERTH
BASIN
Continent-ocean boundary
INDIAN
OCEAN
PERTH
AFO
Figure 3
EUCLA BASIN
BIGHT
BASIN
EYRE SUB-BASIN
RECERCHE SUB-BASIN
Esperance
Albany
BREMER BASIN
28°
32°
36°
114°
120°
126°
SOUTHERN
OCEAN
400 km
N
AFO Albany-Fraser Orogen
PO Pinjarra Orogen
Major Mesozoic depocentre
ODP borehole (Leg 182)
Salt lake (playa)
Primary valley with active drainage
Primary valley with inactive drainage
Major drainage divide
Jarrahwood axis

The purpose of this article is to describe the inset-valleys in terms of their spatial and geometric properties, estimate their age of incision, and then infer which allogenic control, or combination of allogenic controls, caused them to form. This latter aspect is particularly problematic because, as is typical of unconformities, buried valleys (i.e., "paleovalleys") are inherently information poor in that very little historical record of their development is preserved. The greater difficulty in interpreting the origin of unconformities compared with the rocks they bound has made them an unpopular choice of study and largely to be overlooked in paleoenvironmental reconstructions in continental settings, despite their pivotal importance in the geological record (Sloss 1963; Schumm and Ethridge 1994; Shanley and McCabe 1994). Perhaps nowhere does this apply more than in low-relief, tectonically stable, and long-term emergent crustal blocks, such as the Yilgarn Craton, where the sedimentary cover is generally thin, poorly preserved, and segmented by numerous unconformities, some of which may have great lateral, temporal, and genetic significance.

Regional Geological and Geomorphic Setting

Geology. The Yilgarn Craton is the largest and oldest of Western Australia's major tectonostratigraphic units (fig. 1) and dominantly comprises NNW-trending belts of strongly metamorphosed and deformed Archean sedimentary and felsic-mafic-ultramafic volcanic and intrusive rocks (greenstones) set within larger areas of weakly metamorphosed and deformed Archean granite (Myers 1993). Phanerozoic sedimentary basins surround the craton except to the north and south, where it is bounded by Proterozoic metamorphic rocks of the Capricorn and Albany-Fraser orogens, respectively (fig. 1).

Most significant of the pericratonic basins with respect to this study are the Eucla and adjoining Bight Basins (fig. 1). The Eucla Basin is a downwarp of Precambrian basement, which, in its western part, contains up to 150 m of Lower–Middle Cretaceous continental to shallow marine mostly fine-grained sediments of the Loongana Sandstone and Madura and Toondi Formations, overlain by up to 30 m of Middle Eocene shallow marine sands of the Hampton Sandstone that are, in turn, overlain by up to 80 m of Middle Eocene–Middle Miocene mostly cool-water carbonates of the Wilson Bluff, Toolinna, and Nullarbor Limestones (Lowry 1970; Hocking 1990; Jones 1990; see fig. 4). The adjoining Bight Basin, situated beneath the continental shelf and slope, was produced by rifting between Australia and Antarctica during the Mesozoic and is filled with up to 12 km of dominantly Upper Jurassic–Upper Cretaceous continental to shallow marine siltstones, sandstones, and shales subdivided into a variety of lithostratigraphic and sequence-stratigraphic units (Bein and Taylor 1981; Hocking 1990; Totterdell et al. 2000). A similar geological setting occurs along the western margin of the Yilgarn Craton, where up to 15 km of primarily continental siliciclastic sediments infill the Perth Basin, produced by rifting between Australia and Greater India in the Early Permian–Early Cretaceous (Harris 1994).

Overlying Precambrian crystalline basement in the eastern Yilgarn Craton are scattered remnants of the Gondwanan Permo-Carboniferous glaciation (Crowell and Frakes 1971). These commonly comprise bouldery diamictites set within shallow basins and a poorly integrated network of narrow, deep (>80 m) "tunnel valleys" cut into bedrock below the late Paleozoic ice sheet (Eyles and de Broekert 2001; fig. 2). Next youngest in the Phanerozoic sedimentary succession are Eocene–?Miocene continental to shallow marine sediments hosted by the inset-valleys (Clarke 1993; de Broekert 2002). In interior parts of eastern Yilgarn Craton, situated beyond the influence of several major marine transgressions in the Middle and Late Eocene (McGowran 1989), the inset-valley fills are composed of a basal coarse-grained fluviatile unit of Middle Eocene age termed the "Wollubar Sandstone," unconformably overlain by a clay-rich paludal unit of probable Oligo-Miocene age termed the "Perkolilli

Figure 1. Distribution of valley systems in southwestern Australia. *A*, Regional location map. *B*, Approximate extent of primary valleys with inset-valleys hosted within their bedrock surface. *C*, Detail of southwestern Australia showing Archean Yilgarn Craton surrounded by Proterozoic orogenic belts and Phanerozoic sedimentary basins. Primary valleys on Yilgarn Craton slope gently away from major NS drainage divide toward the Perth and Bight rift basins, indicating that the form of their bedrock surface was established by a major Mesozoic drainage system. *WA* = Western Australia; *SA* = South Australia; *NT* = Northern Territory. Adapted from Cope (1975), van de Graaff et al. (1977), and Alley and Beecroft (1993).

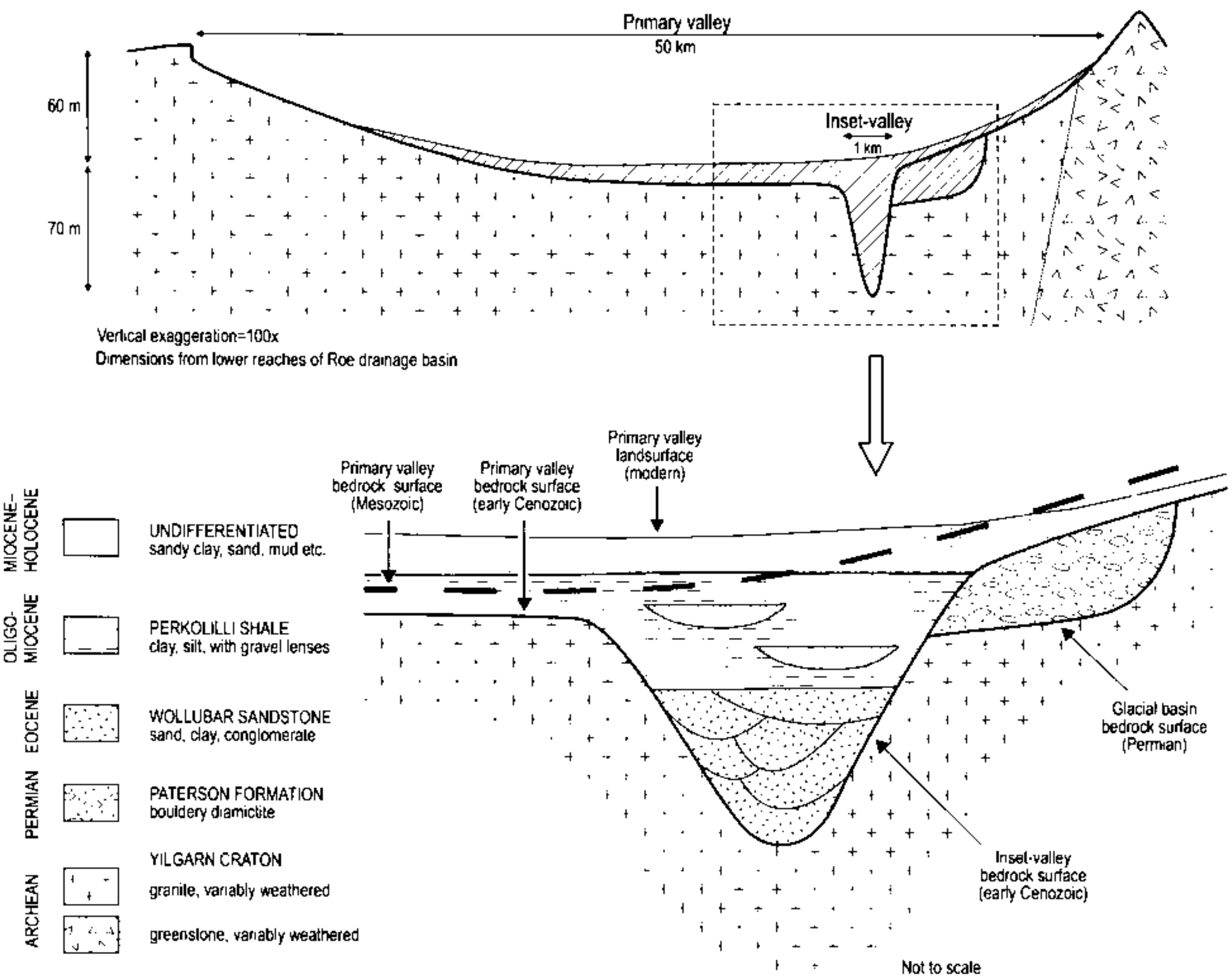


Figure 2. Main geomorphic and stratigraphic components of a typical valley in the eastern Yilgarn Craton. With decreasing age, the major phases of valley incision and filling are (1) glacial incision and filling of tunnel valleys and basins during the Permian; (2) fluvial incision and partial filling of primary valleys during the Mesozoic (dashed line shows inferred Mesozoic bedrock surface); (3) fluvial incision of inset-valleys and stripping of Mesozoic sediment and a small thickness of weathered Precambrian bedrock from primary valleys during the early Cenozoic; (4) filling of inset-valleys during the early-mid-Cenozoic; (5) partial filling of primary valleys and burial of inset-valleys and their fills during the late Cenozoic. Note that although the age of the primary valley bedrock surface is early Cenozoic (being part of the same regional erosive surface as the inset-valleys), its form is largely inherited from a valley system created in the Mesozoic. Upper diagram has very high vertical exaggeration (~ × 100).

Shale" (Kern and Commander 1993; fig. 2). The inset-valleys and their fills are, in turn, unconformably overlain by a sequence of Miocene-Holocene ephemeral-fluvial, playa-lake, playa, and aeolian sediments (Glassford 1987; Clarke 1993), the thickness and complexity of which increases to the north of the eastern Yilgarn Craton.

Much of the Precambrian crystalline basement and its Phanerozoic sedimentary cover are deeply weathered, in the former instance typically consisting of a thick kaolinite-rich saprolite locally overlain by a thin ferruginous, pisolitic to nodular "duricrust" (Anand and Paine 2002). The mineralogical and textural composition of the inset-valley fills indicates that major phases of weathering occurred before the inset-valleys were incised as well as during and after they were filled (de Broekert 2002).

Geomorphology. The landsurface of Yilgarn Craton forms part of the Great Plateau of Western Australia—a vast terrain of gently undulating and subdued relief (Jutson 1934). Within the central Yilgarn Craton portion of the Great Plateau, termed the "Yilgarn Plateau" by Jennings and Mabbutt (1986), the landsurface elevation rarely exceeds 650 m and is mostly between 250 and 450 m above mean sea level. A major NS-trending continental drainage divide bisects the Yilgarn Plateau into valleys that slope toward the Indian Ocean in the west and the Southern Ocean in the south (Beard 1973, 1998; fig. 1). These valleys, herein termed "primary valleys," form a well-integrated contributory pattern, are typically subrectangular to rectangular in shape, 20–100 km wide, and are of very low gradient (0.04%–0.008%) and relief (50–150 m; fig. 2). Owing to the region's semiarid to arid climate, drainage within the valleys is currently internal, with ill-defined ephemeral streams terminating in extensive flats or salt lakes (playas) along the valley bottoms (fig. 1). Outcrops of variably weathered bedrock along the upper valley flanks and greater thicknesses of sediment beneath the valley floors indicate that the landsurface form of the primary valleys closely reflects the form of their bedrock surface (fig. 2).

The large-sized contributory pattern and close spatial association of the primary valleys with the Eucla-Bight and Perth basins (fig. 1) suggest that the form of their bedrock surface was largely established by fluvial erosion during the Mesozoic (Beard 1973, 1998, 1999; Johnstone et al. 1973; Bunting et al. 1974; van de Graaff et al. 1977; Clarke 1994*b*). However, the absence of any known Mesozoic sediment within the primary valleys, except for in rare, small, fault-bounded basins in the western Yilgarn Craton (Le Blanc Smith 1993), indicates that subsequent incision of the inset-valleys was accompanied by the generation of a regional unconformity involving widespread stripping of sediment and perhaps also a small thickness of bedrock well beyond the inset-valley flanks (de Broekert 2002). Consequently, the bedrock surface of the primary valleys is either directly overlain by late Cenozoic arid-zone sediments or, less commonly, by Tertiary sediments deposited in excess of inset-valley accommodation (fig. 2).

Nomenclature

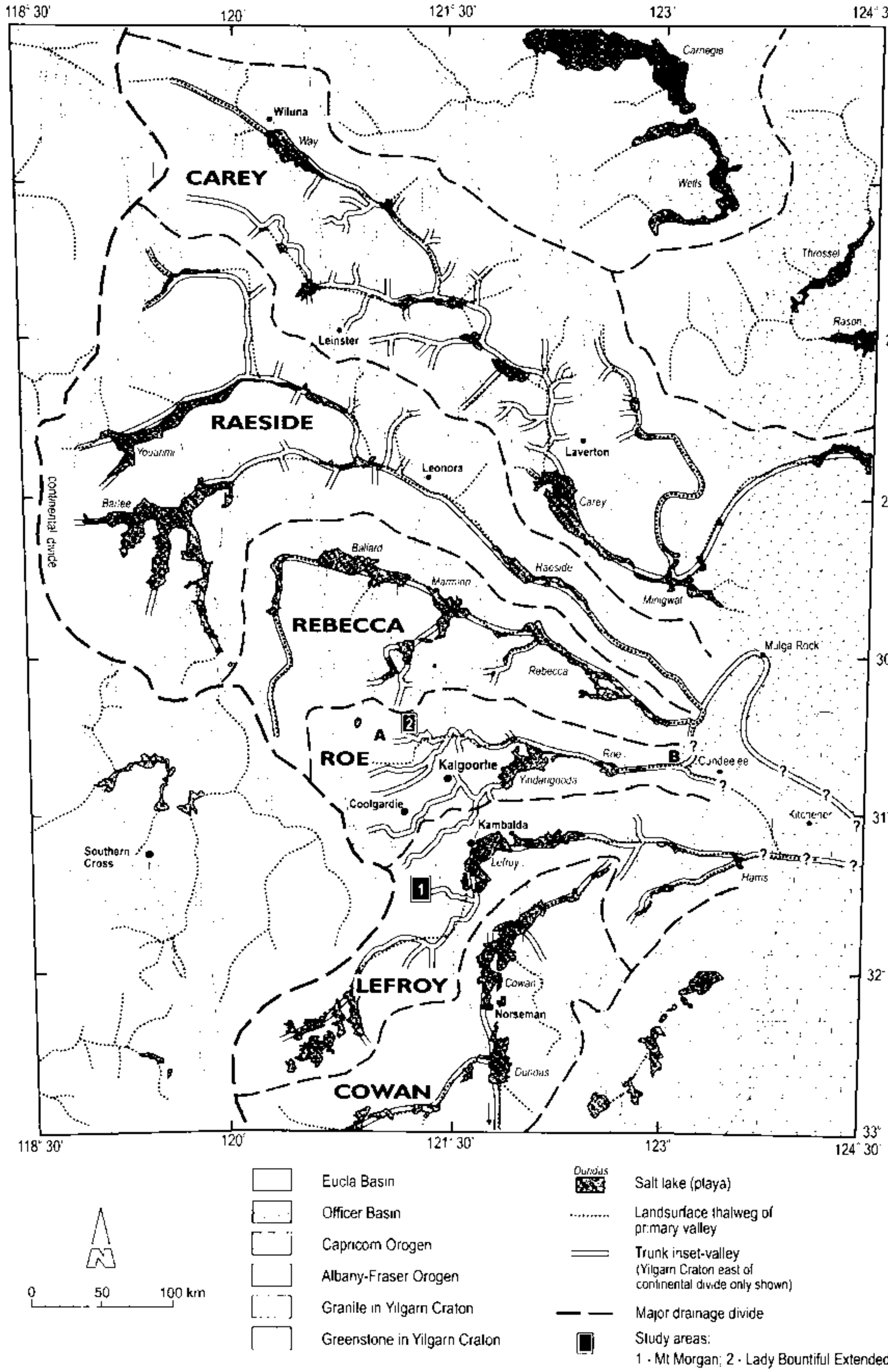
The term "inset-valley," as opposed to "paleochannel" (e.g., Kern and Commander 1993; Anand and Paine 2002), is justified on the basis that valleys are landforms that host river systems, whereas channels are landforms within river systems (Bates and Jackson 1987). Consequently, valleys and channels are vastly different in their dimensions. For example, the channel of the Mississippi River (the world's fourth largest) is about 600 m wide and 18 m deep at Vicksburg, some 500 km upstream from its delta (Schumm and Ethridge 1994), which is about the same size as a typical second-order (sensu Strahler 1952) inset-valley draining a small headwater catchment in the eastern Yilgarn Craton. A major difference between valleys and channels also exists in their style of fill, with valleys hosting a much greater variety of sedimentary deposits, commonly including fluvial channel fills, as indeed occurs in the case of the inset-valleys (fig. 2; de Broekert 2002). Similarly, the term "incised-valley" (Zaitlin et al. 1994) is avoided because it does not convey the key hierarchical and spatial context of a valley within a valley, and except for those produced by tectonism or dissolution, most valleys are "incised."

Geomorphology of the Inset-Valleys

Spatial Distribution of Inset-Valleys. Figure 3 shows the distribution of "trunk" (greater than or equal to fifth-order) inset-valleys in the eastern Yilgarn Craton as mapped by regional groundwater and mineral exploration drilling programs (Smyth and Button 1989; Fulwood and Barwick 1990; Kern and Commander 1993; Johnson et al. 1999), and the landsurface thalwegs of the primary valleys as mapped from the distribution of salt lakes, aerial photography, surficial geology, and topographic data (Beard 1973; Bunting et al. 1974; van de Graaff et al 1977). A close spatial association between the two is evident, which, given the similarity in form of the landsurfaces and bedrock surfaces of the primary valleys, clearly indicates that incision of the inset-valleys was strongly controlled by the ancient landsurface provided by the bedrock surface of the primary valleys.

Owing to a lack of subsurface information, it is difficult to establish whether the inset-valleys extend from the eastern Yilgarn Craton and Albany-Fraser Orogen into the Eucla Basin (fig. 3). However, dendritic patterns in isopach maps of the Hampton Sandstone NW of Kitchener (Griffin Coal Mining Company, unpub. data, 1981), which fills the base of the trunk inset-valley near Lake Harris (Jones 1990), suggests that this occurs. In this setting, the inset-valleys are likely to be incised within the top of the Cretaceous Madura and Toondi Formations, largely filled with the Middle Eocene Hampton Sandstone and then buried beneath the Middle Eocene–Middle Miocene Wilson Bluff, Toolinna, and Nullarbor Limestones (fig. 4).

Buried inset-valleys containing palynologically dated or lithostratigraphically correlated Eocene sediments at their base also occur west of the continental divide on the Yilgarn Craton (Smyth and


118° 30'
120°
121° 30'
123°
124° 30'
26°
27°
28°
29°
30°
31°
32°
33°
Carnegie
Wiluna
Way
CAREY
Wells
Throssel
Rason
Leinster
RAESIDE
Youanmi
Laverton
Leonora
Barlee
continental divide
Carey
Ballard
Raeside
Marmion
Minigwal
REBECCA
Mulga Rock
Rebecca
A
2
ROE
Kalgoorlie
Roe
B
Cundeelee
Yindarlgooda
Coolgardie
Kitchener
Kambalda
Southern Cross
Lefroy
Harris
1
LEFROY
Cowan
Norseman
Dundas
COWAN
Eucla Basin
Officer Basin
Capricorn Orogen
Albany-Fraser Orogen
Granite in Yilgarn Craton
Greenstone in Yilgarn Craton
Dundas
Salt lake (playa)
Landsurface thalweg of primary valley
Trunk inset-valley
(Yilgarn Craton east of continental divide only shown)
Major drainage divide
Study areas:
1 - Mt Morgan; 2 - Lady Bountiful Extended
0
50
100 km

Button 1989; Waterhouse et al. 1994; Robertson et al. 2001) and Capricorn Orogen (Robertson et al. 1999), along the south coast within the onshore Bremer Basin (Comet Resources NL, unpub. data, 1999), and around the northeastern margin of the Eucla Basin in South Australia (Benbow et al. 1995; Alley et al. 1999; Hou et al. 2003*a*, 2003*b*; fig. 1). Although separate outlets to the sea prevent lateral continuity between these and the inset-valleys in the eastern Yilgarn Craton from being unequivocally established, it is likely that they all form part of the same valley system. With a combined aerial extent of $\sim 1.2 \times 10^6$ km^2 (excluding a further 0.2×10^6 km^2 possibly lying beneath Eocene–Miocene limestones in the onshore Eucla Basin), this makes the inset-valleys among the largest, if not the largest, of paleovalley systems known globally. A comparably large network of paleovalleys occurs along the Cretaceous Western Interior Seaway in North America, where individual buried valley systems have been traced for up to 320 km inland from the paleoshoreline and over a region of 85,000 km^2 (Zaitlin et al. 2002; Plint and Wadsworth 2003).

Drainage Basin Area and Shape of Inset-Valleys. The major drainage basins for the primary valleys and hence the inset-valleys in the eastern Yilgarn Craton are shown in figure 3. Drainage basin areas range from about 17,000 km^2 for the Roe Basin to about 65,000 km^2 for the Raeside Basin. Detailed subsurface mapping of the inset-valley network in the Kalgoorlie region (Kern and Commander 1993; de Broekert 2002) indicates that the Roe drainage basin contains about seven orders of inset-valleys and is therefore a seventh-order drainage basin. Much larger drainage basins, such as the Raeside and Carey, are probably one or two orders higher.

Except for the upper parts of the Raeside and Rebecca, the drainage basins are approximately rectangular in shape and cut across the lithological and structural grain of the Precambrian bedrock (fig. 3). The Carey, Raeside, Rebecca, Roe, and Lefroy drainage basins are bounded by the continental divide in their headward tracts and form a radial pattern focused on an area within the Eucla Basin to the southeast of Cundeelee. This pattern is accentuated if the divide between the Cowan and Lefroy primary valleys is removed, which is regarded by Clarke (1994*a*) to have developed in the Jurassic because of downward tilting of the craton margin toward the evolving Bight Basin. The radial pattern in drainage basins probably resulted from late Mesozoic subsidence within the Eucla Basin and the Albany-Fraser Orogen having imposed a barrier to the Mesozoic rivers that was breached in the vicinity of Lake Harris and Cundeelee (fig. 3).

Pattern of Inset-Valleys. The pattern of the inset-valley networks in the eastern Yilgarn Craton is dominantly subdendritic (Howard 1967), exhibiting a small but distinct influence of initial slope and basement lithology and structure. The influence of initial slope is obvious in that the trunk inset-valleys follow the pattern established by the bedrock surface of the primary valleys. Perhaps the most outstanding example of structural control occurs SW of Mulga Rock, where the combined Rebecca-Raeside trunk inset-valley follows the contact between Precambrian basement of the Yilgarn Craton and Permian sediments of the Officer Basin for about 100 km before making a broad U-turn back to the Eucla Basin (Fulwood and Barwick 1990; fig. 3). Pronounced structural control is also evident in the Kalgoorlie area, where many trunk inset-valleys have a NE orientation paralleling a regional fracture direction (Johnson and McQueen 2001), and in the Mt. Morgan area, where a trunk inset-valley preferentially follows weakly resistant Archean metasediments and the contact between Archean granite and mafic-ultramafic rocks (fig. 5).

Texture of Inset-Valleys. The texture of a stream or valley network is commonly quantified in terms of drainage density (ratio of total tributary length to drainage basin area), frequency (ratio of total number of tributaries to drainage basin area), and bifurcation ratio (ratio of the total number of tributaries within a given order to that of the next highest order; Horton 1945). For the small (44 km^2), fourth-order Lady Bountiful Extended drainage basin situated in the upper Roe drainage basin (fig. 3), the values of inset-valley density, frequency, and bifurcation ratio are 0.7 km/km^2, 1.0/km, and 3.0, respectively (fig. 6). The bifurcation ratio falls

Figure 3. Distribution of "trunk" (≥fifth-order) inset-valleys in the eastern Yilgarn Craton in relation to landsurface thalwegs of primary valleys. A close correspondence between the two indicates that incision of the inset-valleys was largely controlled by the bedrock surface of the primary valleys. Extension of the inset-valleys into the Eucla Basin is poorly constrained but seems probable. Location of eastern Yilgarn Craton shown in figure 1. Adapted from Beard (1973), Bunting et al. (1974), Kern and Commander (1993), and Johnson et al. (1999).

within the lower range for drainage networks developed in homogenous rocks, but the density and frequency values are low compared with modern drainage basins (Chorley et al. 1984, p. 318–319). In the case of density, this is largely because of the first-order inset-valleys not having been mapped in the subsurface to their point of initiation, and in the case of density and frequency, it may be due to the presence of relatively fresh granite within the eastern part of the drainage basin at the time of inset-valley incision (fig. 6). Assuming that the bifurcation ratio of 3.0 at Lady Bountiful Extended is representative and that the trunk inset-valley draining the Roe Basin is of seventh order, it can be calculated that approximately 2000 first-order inset-valley tributaries occur within the Roe drainage basin (fig. 3).

Transverse Form of Inset-Valleys. Open-cut gold mine exposures (de Broekert 2002) and cross sections constructed from borehole transects (Kern and Commander 1993; Johnson et al. 1999) reveal that the inset-valleys can be divided into five major groups based on their transverse form (fig. 7).

Most of the inset-valleys (group 1) are symmetric and U shaped, comprising narrow to moderately broad concave-up valley floors subtended by gently inclined, straight to sigmoidal side-slopes. A small, first-order inset-valley of this type is shown in figure 8. The other groups comprise inset-valleys with a strongly asymmetric V shape (group 2), two-tier valleys with a narrow "inner" valley or alluvial channel set within a broader "outer" valley (group 3), W-shaped valleys with two floors separated by at least 500 m horizontally and up to 35 m vertically (group 4), and inset-valleys with one or more steps in their side-slopes (group 5).

Borehole transects at Mt. Morgan indicate that asymmetry in the group 2 inset-valleys was dominantly caused by incision along the contact between two bedrock types with markedly different resistance to erosion, in this case, strongly weathered granite and slightly weathered ultramafic rocks (fig. 5). The group 3 inset-valleys are likely to be "inner-channels" formed by local scouring and overdeepening of the valley floor, as observed in modern streams and simulated in flumes (Shepherd and Schumm 1974).

The group 4 inset-valleys, an example of which also occurs at Mt. Morgan (fig. 5), were probably formed by the "formative" stream splitting around a bedrock high, with one of the branches becoming abandoned before incision of the valley was completed. Steps in the side-slopes of the group 5 inset-valleys are, in most cases, likely to be structural benches formed where fluvial incision was temporarily retarded on reaching relatively fresh bedrock (Thornbury 1954, p. 112). In such cases, the steps show no clear relationship with the adjoining fluvial sediments (Wollubar Sandstone), but in others, the steps are directly overlain by gravel lags that extend laterally into the main body of the Wollubar Sandstone and clearly mark the base of fluvial channel fills (fig. 9). Unlike the structural benches, these were therefore formed while the inset-valleys were being filled. Importantly, the steps are in both instances unrelated to bedrock terraces formed by episodes of stability followed by incision, such as would be caused by episodic adjustments in base-level (Thornbury 1954, p. 157–158) or the complex response of fluvial systems to changes in sediment supply (Schumm and Parker 1973).

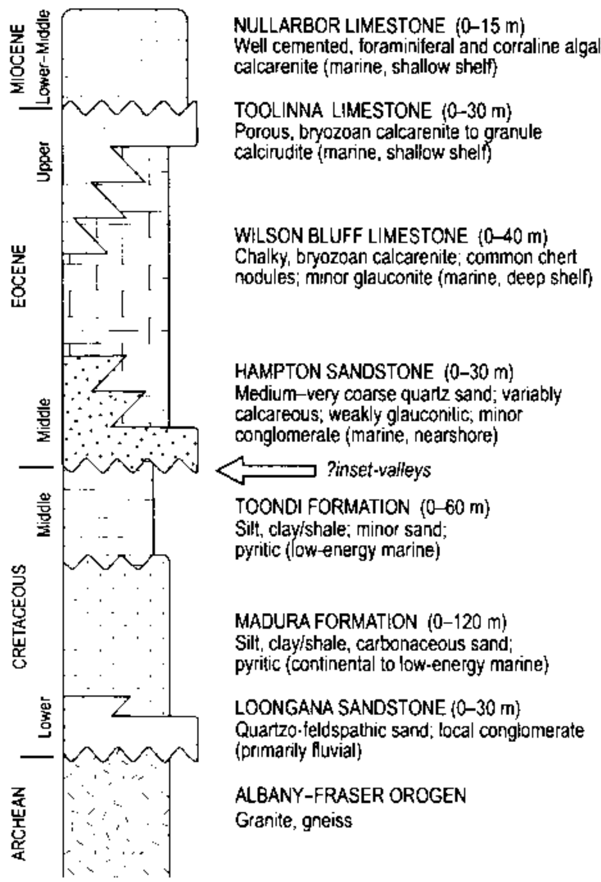


Figure 4. Schematic stratigraphic section of the western Eucla Basin. Arrow indicates probable development of inset-valleys along the Cretaceous-Eocene unconformity. Adapted from Lowry (1970), Hocking (1990), and Jones (1990).

A feature that all of the inset-valleys have in common is that the inclination of their side-slopes is fairly low, averaging about 10% (6°) and reaching

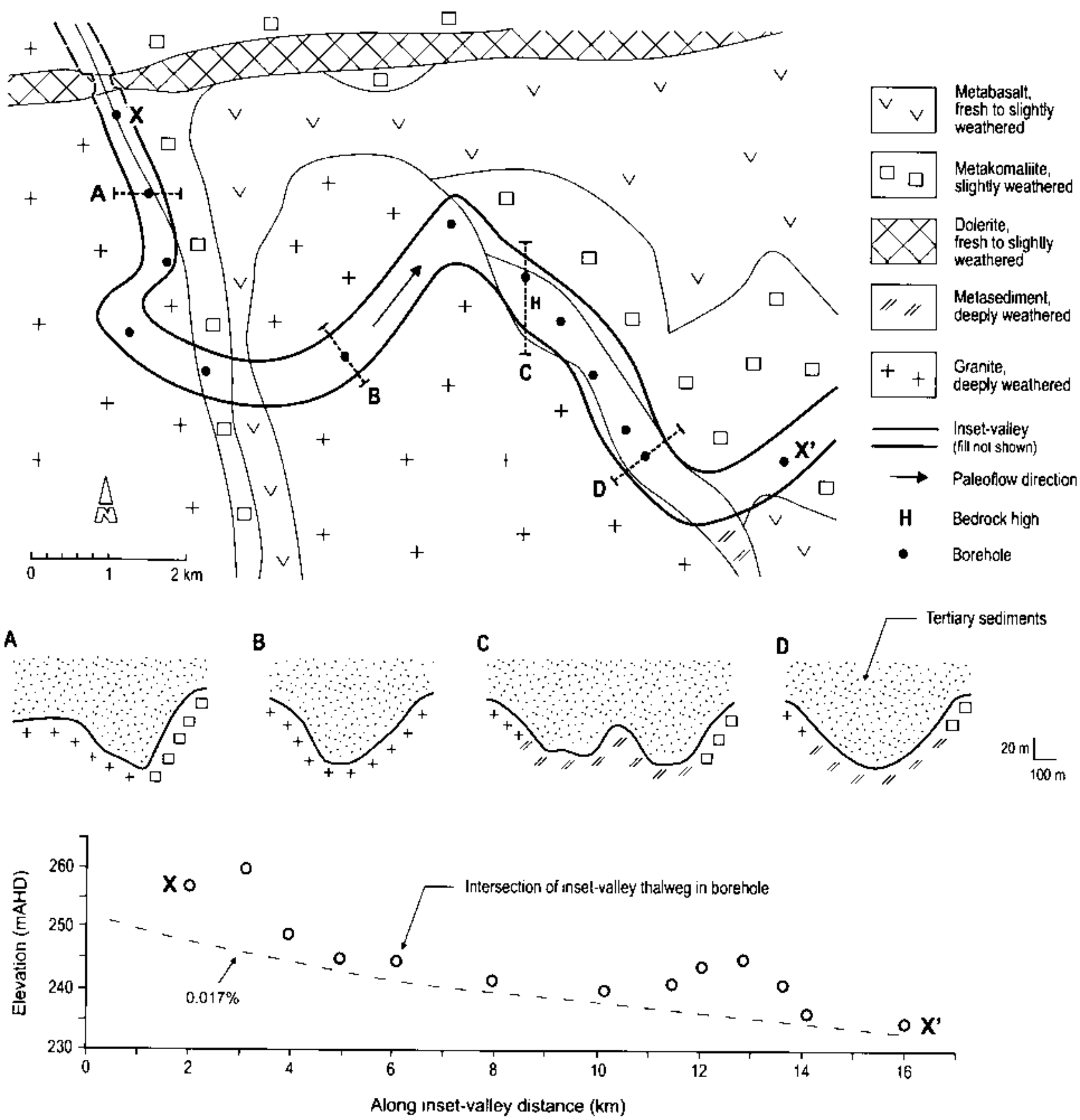


Figure 5. Map of Precambrian basement and trunk inset-valley at Mt. Morgan showing strong lithological and structural control on inset-valley incision (*top*). Longitudinal profile (*bottom*) of inset-valley thalweg shows very gentle overall slope (~0.017%) punctuated by broad, low-relief highs produced by resistant bedrock at the time of inset-valley incision. Variations in bedrock erodibility also reflected by a change in transverse form from typical U shape to W shape or U shape with strong side-slope asymmetry (*middle*). Width of inset-valley in top panel taken at top of fluvial, sandy part of fill (Wollubar Sandstone). Location of Mt. Morgan area shown in figure 3. Note vertical exaggeration of ~ × 5 applied to inset-valley cross sections and vertical exaggeration of ~ × 100 applied to longitudinal profile; *m*(*AHD*) = meters above Australian Height Datum.

a maximum of 40% (22°) for the most '"deeply incised" inset-valleys. This indicates that denudation of the valley sides was easily able to keep pace with vertical incision, which, in turn, suggests that the rate of incision was low and that the valley sides were deeply weathered and easily eroded.

Depth and Width of Inset-Valleys. The depth and width, as measured at half the depth of the inset-valleys, range from about 10 and 60 m, respectively, in first-order inset-valleys at Lady Bountiful Extended, to about 120 m and 2.5 km, respectively, in the highest-order inset-valley within the Carey drainage basin. This increase in dimensions with tributary order is characteristic of stream channels and valleys in humid regions, where discharge increases as it moves down the drainage network (Leopold et al. 1964, p. 241–248). On average, the inset-valleys are about 15 times wider (at half their depth) than they are deep, with the width-to-depth ratio ranging from about 10 to 30 and probably primarily controlled by bedrock erodibility, as indicated by the pronounced narrowing of inset-valleys where they cross a resistant dolerite dike at Lady Bountiful Extended (fig. 6).

In comparison to most other paleovalleys (e.g., Schumm and Ethridge 1994; Thorne 1994; Blum and Price 1998), the inset-valleys are narrow. This suggests that their development was terminated at an early stage by a change in allogenic conditions that caused them to fill and that there was limited time for lateral stream erosion to affect appreciable widening of the valley floors.

Longitudinal Form and Gradient of Inset-Valleys. Figure 10 shows the longitudinal form of the Roe inset-valley network between Zuleika Pit, in which a small headwater tributary is exposed, and where the trunk inset-valley leaves the Roe drainage basin, about 30 km west of Cundeelee (fig. 3). Although some irregularities in the valley floor are detected, these are small, and the overall profile of the bedrock-valley floor is remarkably smooth. Gradients decrease consistently from about 0.33% between Zuleika Pit and borehole KRA8 to about 0.01% east of borehole J10, thereby producing an elongate, concave-up profile. A similar progression of gradients occurs within the remainder of the Roe inset-valley network and within inset-valleys in the Raeside and Carey drainage basins (de Broekert 2002).

Closely spaced borehole transects at Mt. Morgan indicate that the irregularities along the inset-valley floors dominantly occur as bedrock highs with an elevation of about 10–15 m above the regional gradient (fig. 5). These are too widely and inconsistently spaced to be riffle-pool or step-pool sequences (Knighton 1998, p. 193–205), and there is no evidence within the inset-valley fills to suggest that they were produced by tectonic deformation, as has been reported to occur in the lower reaches of inset-valleys in South Australia (Hou et al. 2003*b*). It therefore seems likely the highs were formed where narrow zones of fresh bedrock crossed the inset-valley floors (Schumm and Ethridge 1994), although these are no longer obvious because of postvalley-incision weathering.

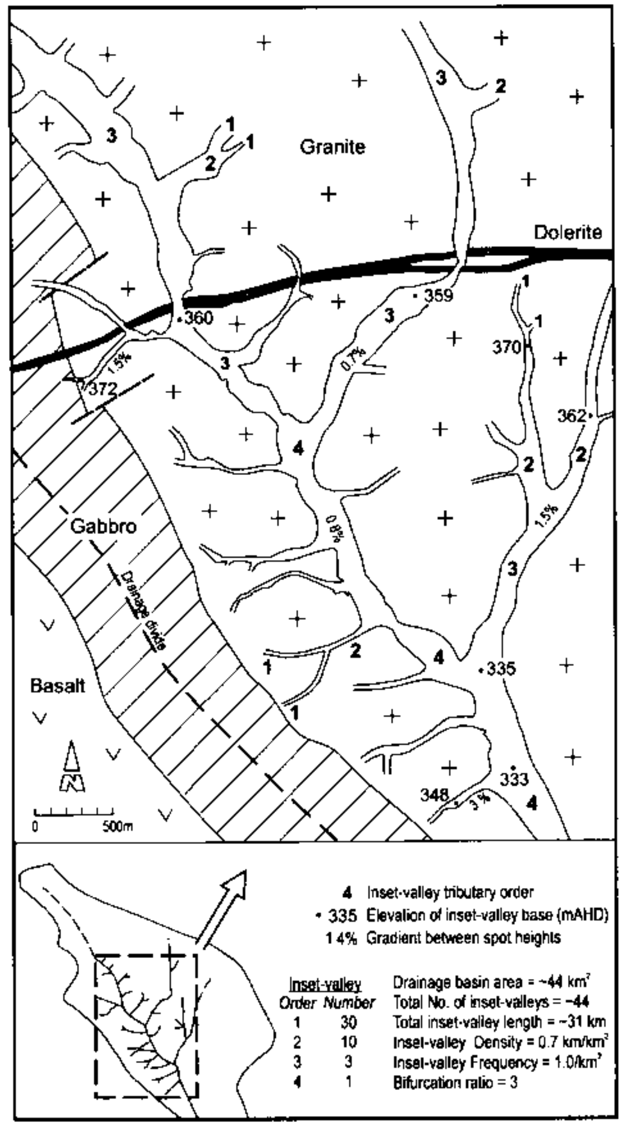
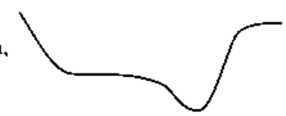


Figure 6. Pattern and width of inset-valleys in the Lady Bountiful Extended drainage basin. Tributary order and longitudinal gradient of inset-valleys also shown. Width of inset-valleys taken at top of fluvial, sandy part of fill (Wollubar Sandstone). Location of Lady Bountiful Extended shown in figure 3; $m(AHD)$ = meters above Australian Height Datum. After 1 : 5000 scale, Tenement Location and Mine Layout Plan, Centaur Mining and Exploration (1995).

Withstanding localized bedrock highs, the overall smoothness and low gradient of the inset-valley longitudinal forms further indicate that variations in bedrock lithology and erodibility were largely masked by deep weathering at the time of inset-

Group	Abundance*	Asymmetry	Morphology and inclination of valley sides	Morphology of valley floor	Example
1	H	None (symmetric) to weak.	Both sides gently inclined and straight to sigmoidal; local minor stepping.	Narrow to broad, open concave-up; occasionally flat.	
2	M	Moderate to strong.	One side steep and straight, other side gentle and sigmoidal with local minor stepping.	Narrow, closed concave-up.	
3	VL	Weak.	Two sets (tiers) of sides, both fairly straight with lower set steeper than upper set.	Narrow, closed V-shaped.	
4	L	Moderate to strong.	Both sides gently to moderately inclined and straight to sigmoidal.	Very broad, weakly to strongly W-shaped, floors >500 m apart laterally & 0-35 m vertically.	
5	L	Moderate to strong.	One slope moderately to steeply inclined and straight to weakly convex-up, other slope with a wide step.	Narrow, closed V-shaped to broad, open, concave-up.	

* H >50%, M 50-30%, L 30-10%, VL <10%

Figure 7. Characteristics of major types of transverse inset-valley form

valley incision and that this took place continuously, as opposed to episodically. It is important to note, however, that incision due to one or more episodes of baselevel fall cannot be discounted on the absence of knickpoints alone because knickpoints developed in massive homogenous material, such as weathered crystalline bedrock, are likely to take the form of rotating headcuts that flatten out as they move upstream (Gardner 1983).

Age of Inset-Valley Incision

There are no means by which the bedrock surface of the inset-valleys can be directly dated. Potentially datable ferruginized (weathered) bedrock occurs directly beneath the flanks of some inset-valleys in the northern Yilgarn Craton (Robertson et al. 1999; Anand and Paine 2002), but it is uncertain whether this formed before, during, or after inset-valley incision. Determination of the age of inset-valley incision is therefore restricted to indirect means.

Palynological analyses of carbonaceous sediments from the base of the inset-valleys indicate that they began to fill in the late Middle–Late Eocene (spore-pollen Middle *Nothofagidites asperus* Zone of Stover and Partridge 1973), situated in the interior of the eastern Yilgarn Craton (Kern and Commander 1993; de Broekert 2002), and perhaps slightly earlier, in the early–middle Middle Eocene (Lower *N. asperus* Zone), situated peripheral to the Eucla Basin in South Australia (Alley and Beecroft 1993; Benbow et al. 1995). Initial filling of the inset-valleys along the inner margin of the Eucla Basin during the Wilson Bluff Transgression (McGowran 1989) with the Hampton Sandstone and lowermost Pidinga Formation (Jones 1990; Hou et al. 2003*b*) indicates a pre–middle Middle Eocene age (>43 Ma) for the inset-valleys in that area. Assuming, as indicated by the geomorphic evidence, that the transition between incision and filling of the inset-valleys was rapid, it therefore seems likely that incision of the inset-valleys terminated in the middle Middle Eocene.

Less reliably dated is the initiation of inset-valley incision. A maximum age of Early Cretaceous (Aptian to Albian) is indicated by incision of the inset-valleys into sediments of that age within the Of-

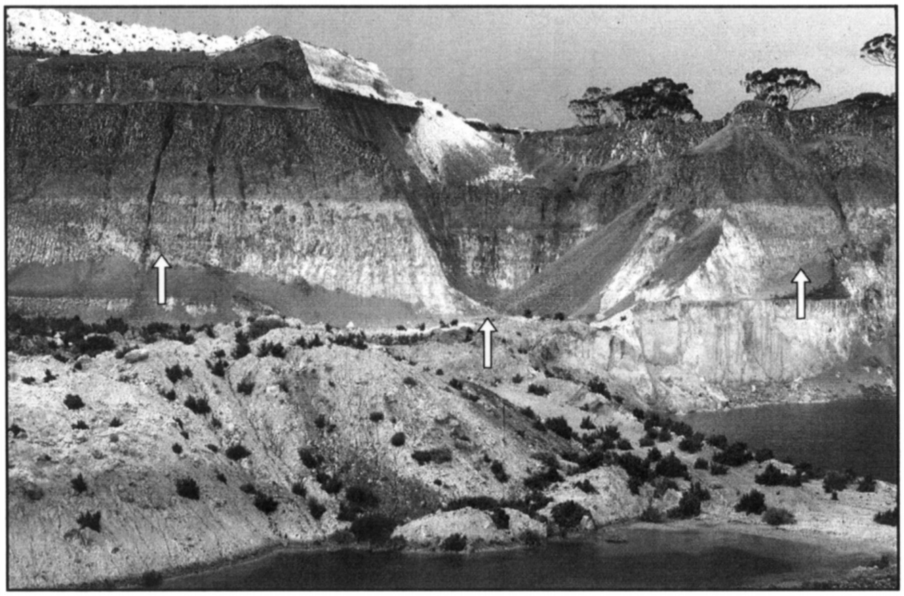


Figure 8. Buried first-order inset-valley (*arrow*) at Lady Bountiful Extended, with symmetric, broad U-shaped form (type 1). Valley depth is ~20 m. Granite bedrock and inset-valley fill are deeply weathered.

ficer and Eucla Basins (Barnes and Pitt 1976; Fulwood and Barwick 1990). Plausibly, an "actual" age for the onset of inset-valley incision could be calculated from the age of cessation of incision using inset-valley depth and an estimate of the rate at which the inset-valleys were incised. In the case of paleovalleys, however, the latter variable is very difficult to quantify, varying widely in response to a complex relationship between the erosive power of the stream (largely a function of discharge and channel width and gradient) and the erodibility of the bedrock (Schumm and Ethridge 1994). Never-

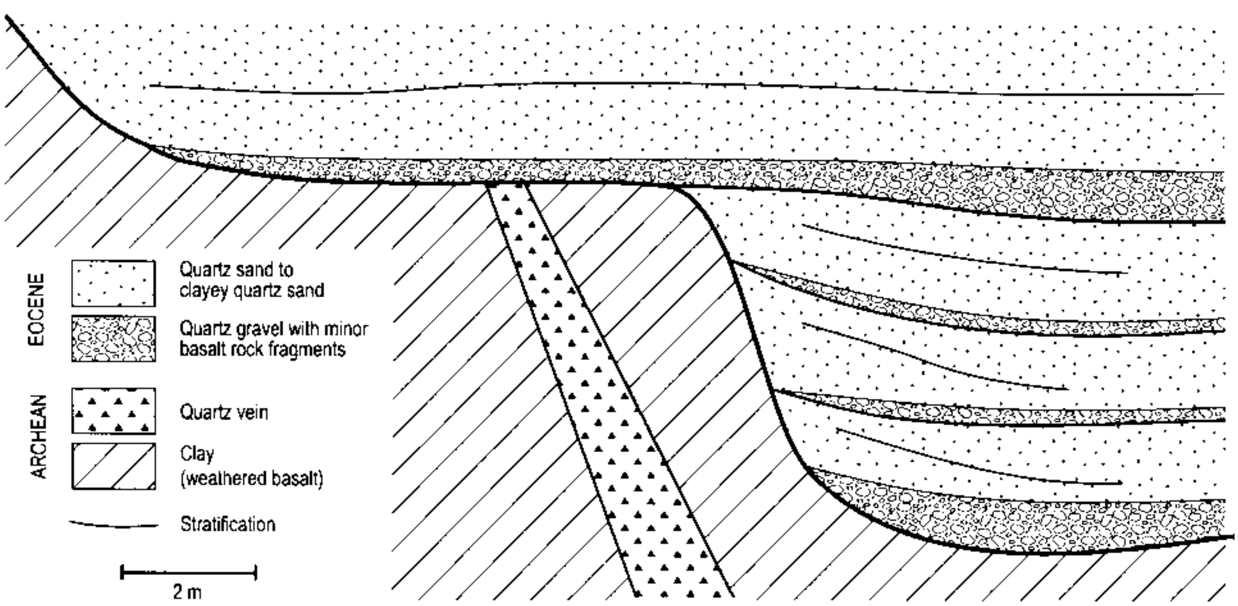


Figure 9. Step in the side-slope of an inset-valley produced by lateral fluvial erosion while the valley was being filled. Basalt bedrock is deeply weathered.

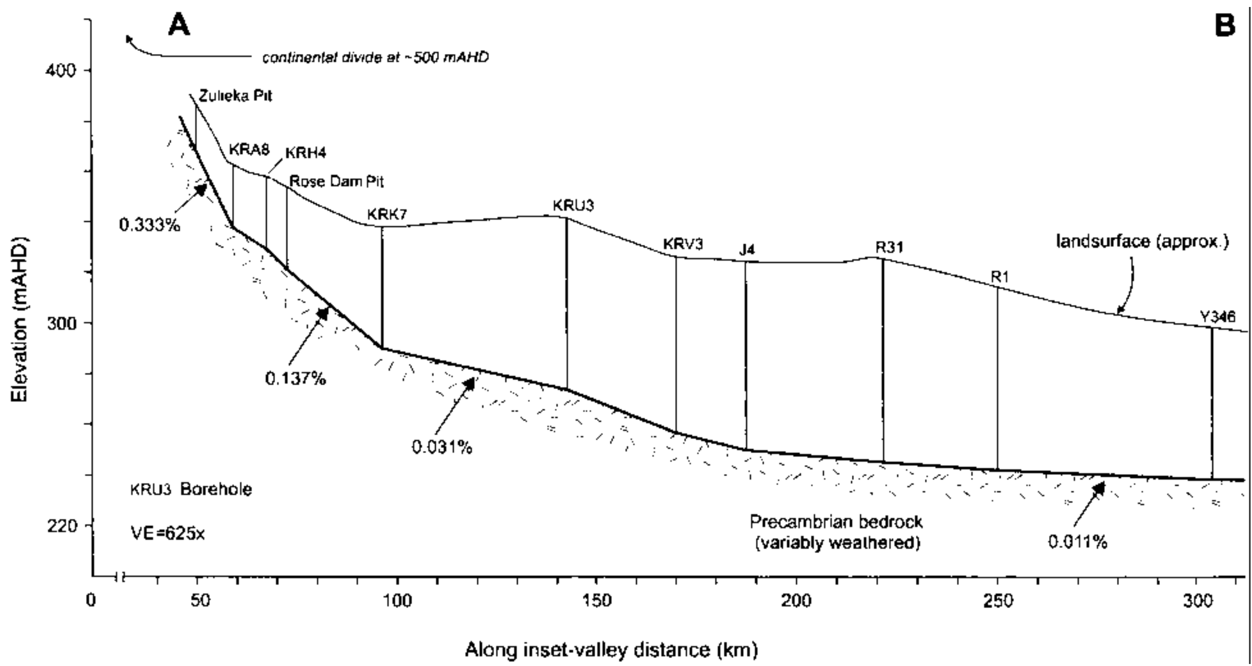


Figure 10. Longitudinal section of the Roe inset-valley network showing smooth concave-up profile with very low gradient in lower reaches. Location of section shown in figure 3; *m(AHD)* = meters above Australian Height Datum.

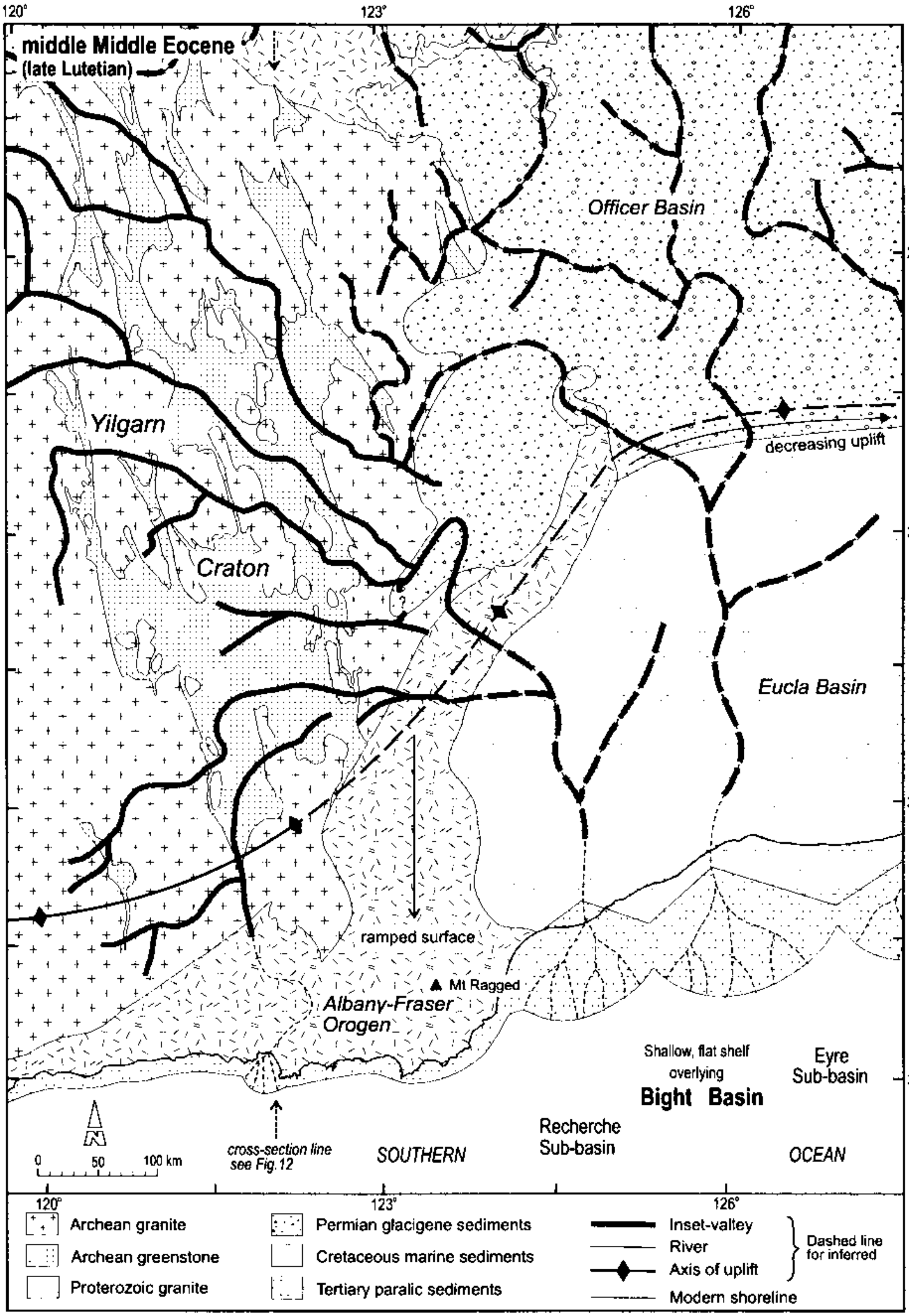
theless, assuming an incision rate of 25 mm/ka, inferred for streams draining the humid, low-relief, and tectonically stable cratonic region of eastern Brazil (Auler et al. 2002), it can be calculated that the inset-valleys took about 5 m.yr. to form, placing their initiation in the earliest Middle Eocene.

The stratigraphic succession within the upper part of the Bight Basin (fig. 1) potentially provides a more direct means of estimating the age during which the inset-valleys were incised, in that valley incision is expected to have resulted in the delivery of a substantial volume of siliciclastic sediment to the shoreline, which, from the latter part of the Late Cretaceous up to deposition of the Hampton Sandstone in the Middle Eocene, remained at or seaward of its present position (Quilty 1994). The only viable contender for such a body of sediment is a prograding wedge of poorly fossiliferous marginal to shallow marine siliciclastic sand recognized in seismic data (Bein and Taylor 1981; Feary and James 1998) and Ocean Drilling Program boreholes (Shipboard Scientific Party 2000; Li et al. 2003) beneath the continental shelf and slope in the vicinity of the Eyre Subbasin (fig. 1). The sand wedge unconformably underlies the Hampton Sandstone (Bein and Taylor 1981) and is probably of early Middle Eocene age (Li et al. 2003), broadly corroborating the estimated age of inset-valley incision.

Control of Inset-Valley Incision

Eustasy. Most readily eliminated as a possible cause of inset-valley incision is a drop in relative sea level (and hence geomorphic baselevel) arising from a lowering of eustatic sea level. Field observations, experimental evidence, and theoretical considerations, reviewed by Schumm (1993), Ethridge et al. (1998), and Blum and Törnqvist (2000), indicate that the fluvial incision resulting from eustasy will be restricted to the coastal plain and newly exposed continental shelf and then only if the shelf is considerably steeper than the coastal plain or the shoreline is lowered onto the continental slope. Lesser fluvial incision may also occur in the region of the paleoshoreline if a wedge of fluvio-deltaic sediment or "coastal prism" was deposited during the previous sea level highstand (Talling 1998).

A major eustatic fall of ~130 m is indicated for the earliest Middle Eocene by the putative "global" sea level curve (Haq et al. 1987), but even if this occurred and brought the shoreline to the shelf edge (best-case scenario for incision), it is highly unlikely that fluvial incision would have propagated more than 250–350 km inland from the lowstand shoreline, which would be the upper limit created by the ~120 m eustatic fall during the Last Glacial Maximum (Saucier 1981). On the basis of the po-



middle Middle Eocene
(late Lutetian)
120°
123°
126°
28°
30°
32°
34°
Officer Basin
Yilgarn
Craton
decreasing uplift
Eucla Basin
ramped surface
Mt Ragged
Albany-Fraser
Orogen
Shallow, flat shelf
overlying
Bight Basin
Eyre
Sub-basin
Recherche
Sub-basin
SOUTHERN
OCEAN
cross-section line
see Fig.12
0
50
100 km
N
Archean granite
Archean greenstone
Proterozoic granite
Permian glacigene sediments
Cretaceous marine sediments
Tertiary paralic sediments
Inset-valley
River
Axis of uplift
Modern shoreline
Dashed line
for inferred

sition of the modern shelf break in the Great Australian Bight, which is probably located shoreward of its Middle Eocene counterpart (Li et al. 2003), this places the inland extent of fluvial incision following any major eustatic fall at the western margin of the Eucla Basin, less than one-third to one-half the distance required to cause incision of the inset-valleys in the headwaters of the Roe, Rebecca, Raeside, and Carey drainage basins (fig. 3).

Also arguing strongly against eustatic fall as being the cause for inset-valley incision is that valleys formed in this way typically have very high width-to-depth ratios (~500 on the basis of data tabulated in Thorne 1994, p. 41) and a depth that rarely exceeds 70 m, even within the zone of maximum incision encompassing the shelf edge or the abandoned coastal prism (Anderson et al. 1996; Blum and Price 1998; Talling 1998; Plint and Wadsworth 2003). This contrasts with the inset-valleys, which typically have a width-to-depth ratio of ~15 and a depth of 100 m over 500 km up-valley from the inferred Middle Eocene shoreline.

An important implication of the inset-valleys, probably not having been caused by a eustatic fall while, at the same time, containing sediments deposited during ensuing eustatic rises (Clarke 1993, 1994*a*), is that the incision and filling of valleys need not be genetically related. Valleys containing marine sediments deposited by a eustatic rise should not, therefore, be attributed to the falling part of a global sea level cycle before the alternative causes for fluvial incision and valley formation have been considered and discounted.

Climate Change. For climate change to affect widespread fluvial incision, it must, in some way, lead to an excess of discharge in relation to sediment supply (Chorley et al. 1984, p. 305–306). Furthermore, the imbalance created must be sufficiently large to exceed the stream's capacity to adjust by changing one or more of its morphological properties, other than gradient, such as sinuosity, width, depth, or bed roughness (see examples of climatically induced stream pattern "metamorphosis" by Schumm [1968] and Baker and Penteado-Orellana [1977]). The confinement of a stream to a narrow valley would also promote vertical incision over other means of stream adjustment, although in the case of the inset-valleys, it is likely that the original streams flowed unrestricted across the broad primary valley floors.

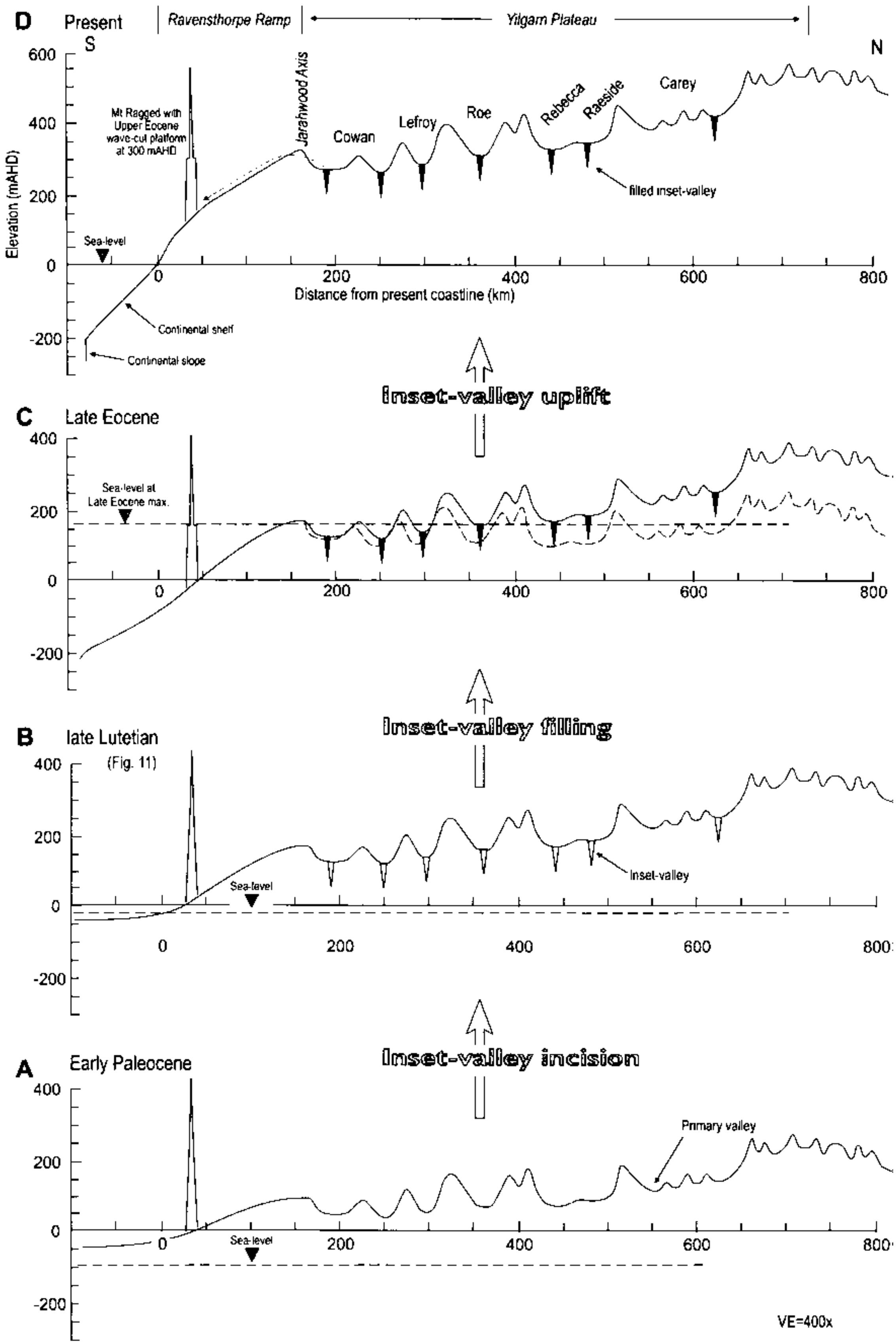
Owing to the large perturbation required, deep and widespread fluvial incision due to climate change is best developed and most evident in valleys glaciated during the Pleistocene. Here, thick valley fills deposited during glacial phases became deeply incised during late glacial and early interglacial times as a result of a marked reduction in sediment supply brought about by decreased glacial outwash (Schumm 1965; Autin et al. 1991) or decreased erosion from hillslopes newly stabilized by vegetation (Vandenberghe 1993; Tebbens et al. 1999). Vandenberghe (1993) suggested that fluvial incision might also occur during the interglacial to glacial transition, when reductions in evapotranspiration result in a marked increase in runoff and fluvial discharge but only a modest increase in sediment supply.

Variations in the other major climatic variable, namely, rainfall, may also induce stream incision. This is particularly the case around the humid-semiarid margin, where small changes in rainfall produce large changes in vegetation cover and also in discharge and sediment supply (Chorley et al. 1984, p. 547). Antevs (1952), for example, attributed the widespread occurrence of arroyos in the semiarid southwestern United States to dramatically increased rates of runoff and, hence, discharge following the removal of vegetation by drought or overgrazing.

A change from a semiarid to a humid climate also has the potential to affect stream incision. Empirical relationships between rainfall and discharge developed by Langbein (1949) and between rainfall and sediment yield developed by Langbein and Schumm (1958), Douglas (1967), and Ohmori (1983) show that the increase in rainfall associated with the change from a semiarid to a humid climate results in an increase in discharge concomitant with a decrease in sediment supply brought about the increase in vegetation cover, thereby creating the opportunity for fluvial incision.

None of the described climatic changes are likely to have been operative in the eastern Yilgarn Craton during the early Middle Eocene, at least to the extent required to have been the main cause for incision of the inset-valleys. Although the onshore

Figure 11. Paleogeographic reconstruction of the eastern Yilgarn Craton and western Eucla Basin during the middle Middle Eocene (late Lutetian). Latitude and longitude are for present geographical location.



D Present
Ravensthorpe Ramp
Yilgarn Plateau
S
N
Mt Ragged with Upper Eocene wave-cut platform at 300 mAHD
Jarahwood Axis
Cowan
Lefroy
Roe
Rebecca
Raeside
Carey
filled inset-valley
Sea-level
Continental shelf
Continental slope
Elevation (mAHD)
Distance from present coastline (km)
Inset-valley uplift
C Late Eocene
Sea-level at Late Eocene max.
Inset-valley filling
B late Lutetian
(Fig. 11)
Sea-level
Inset-valley
Inset-valley incision
A Early Paleocene
Primary valley
Sea-level
VE=400x

stratigraphic and paleobotanical record for the early Middle Eocene along the southern margin of Australia is very poor (McGowran et al. 1997), in keeping with this period of widespread erosion, palynological data and paleogeographic reconstructions for the late Middle Eocene and Late Eocene suggest that a warm, humid climate and cover of nonseasonal, mesothermal (subtropical to warm temperate) rain forest dominated the eastern Yilgarn Craton during the early Middle Eocene (Macphail et al. 1994; Quilty 1994; Alley et al. 1999). Although not in itself conducive to fluvial incision, this setting does however represent a major cooling and less certain drying of climate from the peak in global Cenozoic warmth and humidity that occurred during the Late Paleocene and Early Eocene (Miller et al. 1987; Zachos et al. 2001). The deterioration, which was coeval with a major cooling of oceanic waters (Oberhänsli 1986; Frakes 1997), occurred as a step at the Early/Middle Eocene boundary and may therefore have temporarily resulted in a substantial decrease in evapotranspiration and in an increase in discharge but not to the extent or with the rapidity that occurs at the onset of a glacial phase, which would be expected to trigger deep and widespread fluvial incision.

Climate change also can be largely eliminated as the principal cause for inset-valley incision on geometric grounds. Arroyos and other entrenched streams resulting from reductions in vegetation cover and increased runoff in semiarid regions are, for example, typically no more than 10–15 m deep and characterized by vertical walls and wide, flat floors (Antevs 1952; Schumm et al. 1984), whereas the inset-valleys are mostly more than 30 m deep and U shaped. Furthermore, because no change in baselevel is involved, the depth of incision resulting from climate change must decrease and ultimately reach zero as the shoreline is reached (Posamentier and James 1993). The inset-valleys, by contrast, deepen and then remain approximately stable in depth for at least as far down-valley as the margin of the Eucla Basin (fig. 10).

Tectonics. A fall in relative sea level and geomorphic baselevel resulting from gentle, broadscale (epeirogenic) uplift has long been regarded as the most effective means of inducing fluvial incision, particularly in regions isolated from active mountain building and where the resultant valleys are deep and laterally extensive. Perhaps the most influential proponent of its use was Davis (1899), who invoked regional uniform uplift to initiate his idealized "geographical cycle" of erosion. A striking example of a modern valley produced by epeirogenic uplift is the Grand Canyon in the western Colorado Plateau (Lucchitta 1979), which reaches up to 1600 m in depth and 24 km in width at its top (Thornbury 1954, p. 111). Similarly, large paleovalleys attributed to epeirogenic uplift include the 200–250-km-long, 150–1500-m-deep Wonoka "canyons" and related buried valleys in the late Neoproterozoic Adelaide fold belt of South Australia (Williams and Gostin 2000) and 60–190-m-deep valleys of unknown length in the Upper Cretaceous–Paleocene succession of West Greenland (Dam and Sønderholm 1998). Importantly, these paleovalley systems are associated with regional unconformities and have width-to-depth ratios and side-slope angles that are similar to the inset-valleys.

Why epeirogenic uplift has the potential to cause greater fluvial incision than eustatic fall, which also results in a drop of relative sea level and geomorphic baselevel, remains unclear, but it is probably largely because the magnitude and duration of tectonic uplift are generally much greater (Leeder 1993), the rate of uplift may over short periods be very high (Summerfield 1991, p. 378–379), and tilting, or faulting, associated with uplift may force streams to incise well inland from the coastline (Holbrook and Schumm 1999).

Thus, on the basis of geometric considerations, it would appear that incision of the inset-valleys was dominantly caused by epeirogenic uplift, with climate change possibly having played a minor contributory role. That the eastern Yilgarn Craton has

Figure 12. Cross-sectional paleogeographic reconstruction showing uplift, incision, and filling of the inset-valleys in eastern Yilgarn Craton. *A*, Landsurface before inset-valley incision (Early Paleocene). *B*, Landsurface after epeirogenic uplift and close to completion of inset-valley incision (late Lutetian; see also fig. 11). *C*, Landsurface as in *B* but with inset-valleys partly to wholly filled with shallow marine to continental sediments (Late Eocene). *D*, Landsurface after second phase of uplift, with fluvial incision being restricted by the change to a much drier climate (present). Dashed line in *B* shows effect of removing 0.022% of post-Eocene northward uptilt. Absolute elevations in *A*–*C* based on assumption that Upper Eocene (Tuketja) highstand reached ~160 mAHD (Australian Height Datum). Position of probable Upper Eocene wave-cut platform on Mt. Ragged, projected from 150 km to the east, is after Lowry (1970, p. 156). Position of section shown in figure 11.

been subject to epeirogenic uplift during the Cenozoic is clearly demonstrated by the anomalous elevation and attitude of sediments deposited by marine transgressions in the Late Eocene and Middle Miocene (Jutson 1934; Johnstone et al. 1973; Bunting et al. 1974; Cope 1975; Jones 1990). Particularly useful as a paleoshoreline marker is the Princess Royal Spongolite (Glauert 1926) deposited in shallow (<40 m) water at the highstand of the Tuketja marine transgression in the earliest Late Eocene (Clarke 1994*a*; Gammon et al. 2000). On the basis of the elevation of temporally and environmentally equivalent spicular sediments in eastern South Australia, where there has been considerably less post-Eocene uplift (Hou et al. 2003*b*), it can be inferred that the Tuketja transgression reached no higher than 160 m above present sea level, necessitating at least 140 m of epeirogenic uplift to bring the Princess Royal Spongolite to its present elevation of 300 m at Lake Lefroy (fig. 4). Outcrops of the Princess Royal Spongolite at 325 m at Cundeelee and Mulga Rock (fig. 4) further indicate that the epeirogenic uplift involved uptilting of the eastern margin of the Yilgarn Craton to the north (Jones 1990). A similar northward increase in elevation, corresponding to ~0.022% of tilt, is displayed by the Nullarbor Limestone along the western margin of the Eucla Basin (van de Graaff et al. 1977), suggesting that much of the post-Eocene epeirogenic uplift occurred after the Middle Miocene.

Although more equivocal than the structural relationships previously cited, apatite fission-track data indicate that denudation rates in the eastern Yilgarn Craton increased sharply during the early Cenozoic, reaching a maximum of ~20 m/m.yr. in the latest Eocene (Kohn et al. 2002). No distinct peak in denudation rate is recorded for the early Middle Eocene, but at the very least, the fission-track data indicate that inset-valley incision occurred during a prolonged period of erosion, which is more commensurate with epeirogenic uplift having been the principal cause of inset-valley incision, as opposed to eustatic fall or climate change.

The cause for epeirogenic uplift leading to development of the inset-valleys and tilting of the Eocene and Miocene marine sediments in the Eucla Basin appears to be related to long wavelength surface deflections associated with deep-mantle processes, which will be the subject of a separate article.

Reconstructions of Inset-Valley Development

Figure 11 shows a paleogeographic reconstruction of the eastern Yilgarn Craton and western Eucla Basin for the late Lutetian (~44 Ma) during the final stages of inset-valley incision and before the onset of the Wilson Bluff transgression in the latest Lutetian (~42.5 Ma). The reconstruction has the following major elements:

1. The continental shelf extends well to the south of its current position, tectonic subsidence, and formation of the modern continental slope begun in the middle Bartonian (~39 Ma; Totterdell et al. 2000; Li et al. 2003).
2. The regional shoreline is seaward of its current position. Supporting evidence for this comes from the Albany region (fig. 1), where Middle–Upper Eocene continental sediments infill an irregular basement topography eroded to ~50 m below present sea level (Hos 1975), and offshore from Esperance, where there are numerous submerged shorelines and benches cut into the continental shelf (Morgan and Peers 1973).
3. Increased sediment supply resulting from inset-valley incision within the base of the primary valleys would have led to the construction of deltas and localized progradation of the shoreline. The volume of material delivered to the shoreline would, however, only have been a fraction of that delivered to the Bight Basin by the primary valleys during the Mesozoic.
4. Uplift leading to incision of the inset-valleys probably took place north of an axis that followed the south coast in the west and then deviated inland around the western margin of the Eucla Basin in the east. On the southern Yilgarn Craton and Albany-Fraser Orogen, the axis of uplift and downward ramped surface to its south are prominent in the modern topography (fig. 1) and have been termed the "Jarrahwood Axis" and "Ravensthorpe Ramp," respectively (Cope 1975). These features may initially have been established in the Mesozoic associated with rifting and formation of the Bight Basin (Clarke 1994*a*), although Clarke places the Jarrahwood Axis further to the north than shown in figure 11.
5. To the south of the axis of uplift and within the Eucla Basin, the inset-valleys would have decreased in depth toward the paleoshoreline. These valleys were probably filled and then buried by the Hampton Sandstone and the Wilson Bluff, Toolinna, and Nullarbor Limestones (fig. 4), deposited during marine transgressions in the latter part of the Eocene and Miocene.

A cross-sectional paleogeographic reconstruction illustrating the main stages in inset-valley development is shown in figure 12. Before inset-valley incision, the landsurface stood at a low elevation and was underlain by a mantle of deeply weathered rocks produced by a warm, humid climate and a major reduction in fluvial erosion within the primary valleys since the Mesozoic (fig. 12*A*). Uplift of the Yilgarn Plateau, possibly commencing during the early Cenozoic, continued to a point where stream gradients downslope of the axis of uplift became sufficiently steep to initiate stream rejuvenation and incision of the inset-valleys (figs. 11, 12*B*). Incision propagated inland and may have been promoted by a deterioration of climate at the Early/Middle Eocene boundary. Shortly thereafter, eustatically controlled rises in relative sea level contributed to filling the inset-valleys in the latest Middle and Late Eocene (fig. 12*C*). Following a phase of minor subsidence associated with deposition of the Abrakurrie and Nullarbor Limestones in the Eucla Basin during the Late Oligocene–Middle Miocene (not shown in fig. 12), renewed uplift then brought the eastern Yilgarn Craton and its Tertiary sedimentary cover to its present elevation (fig. 12*D*).

The change from a humid to a dominantly dry climate in the Middle Miocene (van de Graaff et al. 1977) or, possibly, earlier in the earliest Oligocene (de Broekert 2002) protected the inset-valleys and their fills from significant fluvial erosion during uplift in the late Neogene (i.e., between fig. 12*C*, 12*D*). Late Neogene northward uptilting of the western margin of the Eucla Basin and eastern margin of the Yilgarn Craton (as indicated by the southward decline in elevation of the Princess Royal Spongolite and Nullarbor Limestone) evidently did not extend for any great distance to the west because this would have led to submergence of the lower to middle reaches of the northern primary valley networks (see dashed line in fig. 12*C*) of which there is no evidence in the stratigraphic record.

Summary and Conclusions

The inset-valleys provide a rich source of information for interpreting the Cenozoic geological history of the eastern Yilgarn Craton and thereby demonstrate the great value to be gained from the study of regional unconformities in the stratigraphic record, particularly where the sediments are thin, altered, and fragmentary.

On the basis of a detailed documentation of inset-valley distribution and morphology, a critical evaluation of the major causes for valley incision and an assessment of regional Cenozoic structural and stratigraphic relationships, it is concluded that incision of the inset-valleys was principally caused by epeirogenic uplift during the early Middle Eocene. The contributory effects of a fall in eustatic sea level on inset-valley incision during this period would have been confined to the coastal plain and newly exposed continental shelf, but the contributory effects of a change in climate at the Early/Middle Eocene boundary may have extended much farther inland. Climate also indirectly assisted inset-valley formation by promoting deep weathering of the Precambrian bedrock within the primary valleys during the Paleocene and Early Eocene.

Apart from providing significant new insights into the geological history of the eastern Yilgarn Craton and adjoining sedimentary basins during the early Cenozoic, the results of this study provide a basis for future investigations of the inset-valley fills, both in terms of their paleoenvironmental significance and as sources of minerals and groundwater. Particularly prospective for resource exploration are the lower reaches of the inset-valleys, which before this study were predicted not to extend for any great distance beyond the margin of eastern Yilgarn Craton.

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REFERENCES CITED

Alley, N. F., and Beecroft, A. 1993. Foraminiferal and palynological evidence from the Pidinga Formation and its bearing on Eocene sea level events and palaeochannel activity, eastern Eucla Basin, South Australia. Mem. Assoc. Australas. Palaeontol. 15:375–393.

Alley, N. F.; Clarke, J. D. A.; Macphail, M.; and Truswell, E. M. 1999. Sedimentary infillings and development of major Tertiary palaeodrainage systems of south-central Australia. *In* Thiry, M., and Simon-Coincon, R., eds. Palaeoweathering, palaeosurfaces and related continental deposits. Int. Assoc. Sedimentol. Spec. Publ. 27:337–366.

Anand, R. R., and Paine, M. D. 2002. Regolith geology of the Yilgarn Craton, Western Australia: implications for exploration. Aust. J. Earth Sci. 49:3–162.

Anderson, J. B.; Abdulah, K.; Sarzalejo, S.; Siringan, F.; and Thomas, M. A. 1996. Late Quaternary sedimentation and high-resolution sequence stratigraphy of the east Texas shelf. *In* De Batist, M., and Jacobs, P., eds. Geology of siliciclastic shelf seas. Geol. Soc. Lond. Spec. Publ. 117:95–124.

Antevs, E. 1952. Arroyo-cutting and filling. J. Geol. 60: 375–385.

Auler, A. S.; Smart, P. L.; Tarling, D. H.; and Farrant, A. R. 2002. Fluvial incision rates derived from magnetostratigraphy of cave sediments in the cratonic area of eastern Brazil. Z. Geomorphol. N.F. 46:391–403.

Autin, W. J.; Burns, S. F.; Miller, B. J.; Saucier, R. T.; and Snead, J. I. 1991. Quaternary geology of the lower Mississippi Valley. *In* Morrison, R. B., ed. Quaternary nonglacial geology: conterminous U.S. Geol. Soc. Am. K-2:547–582.

Baker, V. R., and Penteado-Orellana, M. M. 1977. Adjustment to Quaternary climatic change by the Colorado River in central Texas. J. Geol. 85:395–422.

Balme, B. E., and Churchill, D. M. 1959. Tertiary sediments at Coolgardie, Western Australia. J. R. Soc. West. Aust. 42:37–43.

Barnes, L. C., and Pitt, G. M. 1976. The Tallaringa palaeodrainage system. Geol. Surv. South Aust. Q. Geol. Notes 59:7–10.

Bates, R. L., and Jackson, J. A. 1987. Glossary of geology. Alexandria, VA, Am. Geol. Inst., 788 p.

Beard, J. S. 1973. The elucidation of palaeodrainage patterns in Western Australia through vegetation mapping. Vegmap Publ. Occas. Pap. 1, 17 p.

———. 1998. Position and developmental history of the central watershed of the Western Shield, Western Australia. J. R. Soc. West. Aust. 81:157–164.

———. 1999. Evolution of the river systems of the southwest drainage division, Western Australia. J. R. Soc. West. Aust. 82:147–164.

Bein, J., and Taylor, M. L. 1981. The Eyre Sub-basin: recent exploration results. APEA J. 21:91–98.

Benbow, M. C.; Lindsay, J. M.; and Alley, N. F. 1995. Eucla Basin and palaeodrainage. *In* Drexel, J. F., and Preiss, W. V., eds. The geology of South Australia. 2. The Phanerozoic. Geol. Surv. South. Aust. Bull. 54:178–186.

Blatchford, T. 1900. The geology of the North Lead, Kanowna. Geol. Surv. West. Aust. Annu. Prog. Rep. 1899, p. 38–47.

Blum, M. D., and Price, D. 1998. Quaternary alluvial plain construction in response to glacio-eustatic and climatic controls, Texas Gulf Coastal Plain. *In* Shanley, K. W., and McCabe, P. J., eds. Relative role of eustasy, climate, and tectonism in continental rocks. SEPM Spec. Publ. 59:31–48.

Blum, M. D., and Törnqvist, T. E. 2000. Fluvial responses to climate and sea-level change: a review and look forward. Sedimentology 47:2–48.

Bunting, J. A.; van de Graaff, W. J. E.; and Jackson, M. J. 1974. Palaeodrainages and Cainozoic palaeogeography of the eastern Goldfields, Gibson Desert and Great Victoria Desert. Geol. Surv. West. Aust. Annu. Prog. Rep. 1973, p. 45–50.

Chorley, R. J.; Schumm, S. A.; and Sugden, D. E. 1984. Geomorphology. London, Methuen, 605 p.

Clarke, J. D. A. 1993. Stratigraphy of the Lefroy and Cowan palaeodrainages, Western Australia. J. R. Soc. West. Aust. 76:12–23.

———. 1994*a*. Evolution of the Lefroy and Cowan palaeodrainage channels, Western Australia. Aust. J. Earth Sci. 41:55–68.

———. 1994*b*. Geomorphology of the Kambalda region, Western Australia. Aust. J. Earth Sci. 41:229–239.

Cope, R. N. 1975. Tertiary epeirogeny in the southern part of Western Australia. Geol. Surv. West. Aust. Annu. Prog. Rep. 1974, p. 40–46.

Crowell, J. C., and Frakes, L. A. 1971. Late Paleozoic glaciation of Australia. J. Geol. Soc. Aust. 17:115–155.

Dam, G., and Sønderholm, M. 1998. Sedimentological evolution of a fault-controlled early Paleocene incised-valley system, Nuussuaq Basin, west Greenland. *In* Shanley, K. W., and McCabe, P. J., eds. Relative role of eustasy, climate, and tectonism in continental rocks. SEPM Spec. Publ. 59:109–121.

Davis, W. M. 1899. The geographical cycle. Geogr. J. 14: 481–504.

de Broekert, P. P. 2002. Origin of Tertiary inset-valleys and their fills, Kalgoorlie, Western Australia. PhD thesis, Australian National University, Canberra, 388 p.

Douglas, I. 1967. Man, vegetation, and the sediment yield of rivers. Nature 215:925–928.

Ethridge, F. G.; Wood, L. J.; and Schumm, S. A. 1998. Cyclic variables controlling fluvial sequence development: problems and perspectives. *In* Shanley, K. W., and McCabe, P. J., eds. Relative role of eustasy, climate, and tectonism in continental rocks. SEPM Spec. Publ. 59:17–29.

Eyles, N., and de Broekert, P. 2001. Glacial tunnel valleys in the Eastern Goldfields of Western Australia cut below the Late Palaeozoic Pilbara ice sheet. Palaeogeogr. Palaeoclimatol. Palaeoecol. 171:29–40.

Feary, D. A., and James, N. P. 1998. Seismic stratigraphy and geological evolution of the Cenozoic, cool-water Eucla platform, Great Australian Bight. Am. Assoc. Pet. Geol. Bull. 82:792–816.

Frakes, L. A. 1997. Grossplots: a method for estimating the temperature state of the Earth and of Australia, Cretaceous to Middle Miocene. Aust. J. Bot. 45:359–372.

Fulwood, K. E., and Barwick, R. E. 1990. Mulga Rock uranium deposits, Officer Basin. *In* Hughes, F. E., ed. Geology of the mineral deposits of Australia and Papua New Guinea. Australas. Inst. Min. Metall., p. 1621–1623.

Gammon, P. R.; James, N. P.; and Pisera, A. 2000. Eocene spiculites and spongolites in southwestern Australia:

not deep, not polar, but shallow and warm. Geology 28:855–858.

Gardner, T. W. 1983. Experimental study of knickpoint and longitudinal profile evolution in cohesive, homogenous material. Geol. Soc. Am. Bull. 94:664–672.

Glassford, D. K. 1987. Cainozoic stratigraphy of the Yeelirrie area, northeastern Yilgarn Block, Western Australia. J. R. Soc. West. Aust. 70:1–24.

Glauert, L. 1926. A list of Western Australian fossils. Geol. Surv. West. Aust. Bull. (suppl.)88:36–71.

Haq, B. U.; Hardenbol, J.; and Vail, P. R. 1987. Chronology of fluctuating sea levels since the Triassic. Science 235:1156–1167.

Harris, L. B. 1994. Structural and tectonic synthesis for the Perth Basin, Western Australia. J. Pet. Geol. 17: 129–156.

Hocking, R. M. 1990. Eucla Basin. *In* Geology and mineral resources of Western Australia. Geol. Surv. West. Aust. Mem. 3:548–561.

Holbrook, J., and Schumm, S. A. 1999. Geomorphic and sedimentary response of rivers to tectonic deformation: a brief review and critique of a tool for recognizing subtle epeirogenic deformation in modern and ancient settings. Tectonophysics 305:287–306.

Horton, R. E. 1945. Erosional development of streams and their drainage basins: hydrophysical approach to quantitative morphology. Geol. Soc. Am. Bull. 56: 275–370.

Hos, D. 1975. Preliminary investigations of the palynology of the Upper Eocene Werillup Formation, Western Australia. J. R. Soc. West. Aust. 58:1–14.

Hou, B.; Alley, N. F.; Frakes, L. A.; Gammon, P. R.; and Clarke, J. D. A. 2003*a*. Facies and sequence stratigraphy of Eocene palaeovalley fills in the eastern Eucla Basin, South Australia. Sediment. Geol. 163:111–130.

Hou, B.; Frakes, L. A.; Alley, N. F.; and Clarke, J. D. A. 2003*b*. Characteristics and evolution of the Tertiary palaeovalleys in the northwest Gawler Craton, South Australia. Aust. J. Earth Sci. 50:215–230.

Howard, A. D. 1967. Drainage analysis in geologic interpretation: a summation. Am. Assoc. Pet. Geol. Bull. 51:2246–2259.

Jennings, J. N., and Mabbutt, J. A. 1986. Physiographic outlines and regions. *In* Jeans, D. N., ed. Australia: a geography. 1. The natural environment. Sydney, Sydney University Press, p. 80–96.

Johnson, C. B., and McQueen, K. G. 2001. The nature of gold-bearing palaeochannel sediments in the Gidji area north of Kalgoorlie, Western Australia. Quat. Int. 82:51–62.

Johnson, S. L.; Commander, D. P.; and O'Boy, C. A. 1999. Groundwater resources of the Northern Goldfields, Western Australia. Water Rivers Comm. Hydrogeol. Rec. Ser. HG2, 57 p.

Johnstone, M. H.; Lowry, D. C.; and Quilty, P. G. 1973. The geology of southwestern Australia: a review. J. R. Soc. West. Aust. 56:5–15.

Jones, B. G. 1990. Cretaceous and Tertiary sedimentation on the western margin of the Eucla Basin. Aust. J. Earth Sci. 37:317–329.

Jutson, J. T. 1934. The physiography (geomorphology) of Western Australia. Geol. Surv. West. Aust. Bull. 95, 366 p.

Kern, A. M., and Commander, D. P. 1993. Cainozoic stratigraphy in the Roe Palaeodrainage of the Kalgoorlie region, Western Australia. Geol. Surv. West. Aust. Prof. Pap. 34:85–95.

Knighton, D. 1998. Fluvial forms and processes. London, Arnold, 383 p.

Kohn, B. P.; Gleadow, A. J. W.; Brown, R. W.; Gallagher, K.; O'Sullivan, P. B.; and Foster, D. A. 2002. Shaping the Australian crust over the last 300 million years: insights from fission track thermotectonic imaging and denudation studies of key terranes. Aust. J. Earth Sci. 49:697–717.

Langbein, W. B. 1949. Annual runoff in the United States. U.S. Geol. Surv. Circ. 52, 14 p.

Langbein, W. B., and Schumm, S. A. 1958. Yield of sediment in relation to mean annual precipitation. Trans. Am. Geophys. Union 39:1076–1084.

Le Blanc Smith, G. 1993. Geology and Permian coal resources of the Collie Basin, Western Australia. Geol. Surv. West. Aust. Rep. 38, 86 p.

Leeder, M. R. 1993. Tectonic controls upon drainage basin development, river channel migration and alluvial architecture: implications for hydrocarbon reservoir development and characterisation. *In* North, C. P., and Prosser, D. J., eds. Characterisation of fluvial and aeolian reservoirs. Geol. Soc. Lond. Spec. Publ. 73:7–22.

Leopold, L. B.; Wolman, M. G.; and Miller, J. P. 1964. Fluvial processes in geomorphology. San Francisco, W. H. Freeman, 522 p.

Li, Q.; James, N. P.; and McGowran, B. 2003. Middle and Late Eocene Great Australian Bight lithobiostratigraphy and the stepwise evolution of the southern Australian continental margin. Aust. J. Earth Sci. 50:113–128.

Lowry, D. C. 1970. Geology of the Western Australian part of the Eucla Basin. Geol. Surv. West. Aust. Bull. 122, 200 p.

Lucchitta, I. 1979. Late Cenozoic uplift of the southwestern Colorado Plateau and adjacent lower Colorado River region. Tectonophysics 61:63–95.

Macphail, M. K.; Alley, N. F.; Truswell, E. M.; and Sluiter, I. R. K. 1994. Early Tertiary vegetation: evidence from spores and pollen. *In* Hill, R. S., ed. History of the Australian vegetation: Cretaceous to Recent. Cambridge, Cambridge University Press, p. 189–261.

McGowran, B. 1989. The later Eocene transgressions in southern Australia. Alcheringa 13:45–68.

McGowran, B.; Li, Q.; and Moss, G. 1997. The Cenozoic neritic record in southern Australia: the biogeohistorical framework. *In* James, N. P., and Clarke, J. D. A., eds. Cool-water carbonates. SEPM Spec. Publ. 56: 185–204.

Miller, K. G.; Fairbanks, R. G.; and Mountain, G. S. 1987. Tertiary oxygen isotope synthesis, sea level history, and continental margin erosion. Palaeoceanography 2: 1–19.

Morgan, K. H., and Peers, R. 1973. Esperance-Mondrain Island, W. A. Geol. Surv. West. Aust., scale 1 : 250,000.

Myers, J. S. 1993. Precambrian history of the West Australian Craton and adjacent orogens. Annu. Rev. Earth Planet. Sci. 21:453–485.

Oberhänsli, H. 1986. Late Cretaceous–Early Neogene oxygen and carbon isotope record at DSDP sites in the Indian Ocean. Mar. Micropaleontol. 10:91–115.

Ohmori, H. 1983. Erosion rates and their relation to vegetation from the viewpoint of world-wide distribution. Bull. Dept. Geogr. Univ. Tokyo. 15:77–91.

Plint, A. G., and Wadsworth, J. A. 2003. Sedimentology and palaeogeomorphology of four large valley systems incising delta plains, western Canada Foreland Basin: implications for mid-Cretaceous sea-level changes. Sedimentology 50:1147–1186.

Posamentier, H. W., and James, D. P. 1993. An overview of sequence-stratigraphic concepts: uses and abuses. *In* Posamentier, H. W.; Summerhayes, C. P.; Haq, B. U.; and Allen, G. P., eds. Sequence stratigraphy and facies variations. Int. Assoc. Sedimentol. Spec. Publ. 18:3–18.

Quilty, P. G. 1994. The background: 144 million years of Australian palaeoclimate and palaeogeography. *In* Hill, R. S., ed. History of the Australian vegetation: Cretaceous to Recent. Cambridge, Cambridge University Press, p. 14–43.

Robertson, I. D. M.; King, J. D.; and Anand, R. R. 2001. Regolith geology and geochemical exploration around the Stellar and Quasar gold deposits, Mt. Magnet, Western Australia. Geochem. Explor. Environ. Anal. 1:353–364.

Robertson, I. D. M.; Phang, C.; and Munday, T. J. 1999. The regolith geology around the Harmony Gold Deposit, Peak Hill, WA. *In* Taylor, G., and Pain, C., eds. New approaches to an old continent. Perth, Coop. Res. Cent. Land. Evol. Miner. Explor., CSIRO, p. 283–298.

Saucier, R. T. 1981. Current thinking on riverine processes and geologic history as related to human settlement in the southeast. Geosci. Man 22:7–18.

Schumm, S. A. 1965. Quaternary paleohydrology. *In* Wright, H. E., and Frey, D. G., eds. The Quaternary of the United States. Princeton, NJ, Princeton University Press, p. 783–794.

———. 1968. River adjustment to altered hydrologic regimen: Murrumbidgee River and palaeochannels, Australia. U.S. Geol. Surv. Prof. Pap. 498, 65 p.

———. 1993. River response to baselevel change: implications for sequence stratigraphy. J. Geol. 101:279–294.

Schumm, S. A., and Ethridge, F. G. 1994. Origin, evolution and morphology of fluvial valleys. *In* Dalrymple, R. W.; Boyd, R.; and Zaitlin, B. A., eds. Incised-valley systems: origin and sedimentary sequences. SEPM Spec. Publ. 51:11–27.

Schumm, S. A.; Harvey, M. D.; and Watson, C. C. 1984. Incised channels: morphology, dynamics and control. Littleton, CO, Water Resourc. Publ., 200 p.

Schumm, S. A., and Parker, R. S. 1973. Implications of complex response of drainage systems for Quaternary alluvial stratigraphy. Nature 243:99–100.

Shanley, K. W., and McCabe, P. J. 1994. Perspectives on the sequence stratigraphy of continental strata. Am. Assoc. Pet. Geol. Bull. 78:544–568.

Shepherd, R. G., and Schumm, S. A. 1974. Experimental study of river incision. Geol. Soc. Am. Bull. 85:257–268.

Shipboard Scientific Party. 2000. Leg 182 summary. *In* Feary, D. A.; Malone, M. J.; et al., eds. Ocean Drilling Program Init. Rep. 182:1–58.

Sloss, L. L. 1963. Sequences in the cratonic interior of the North America. Geol. Soc. Am. Bull. 74:93–114.

Smyth, E. L., and Button, A. 1989. Gold exploration in the Tertiary palaeodrainage systems of Western Australia: Gold forum on technology and practices "World Gold '89," p. 13–22.

Stover, L. E., and Partridge, A. D. 1973. Tertiary and Late Cretaceous spores and pollen from the Gippsland Basin, Southeastern Australia. Proc. R. Soc. Vic. 85:237–286.

Strahler, A. N. 1952. Dynamic basis for geomorphology. Geol. Soc. Am. Bull. 63:923–938.

Summerfield, M. A. 1991. Global geomorphology. Harlow, Prentice-Hall, 537 p.

Talling, P. J. 1998. How and where do incised valleys form if sea level remains above the shelf edge? Geology 26: 87–90.

Tebbens, L. A.; Veldkamp, A.; Westerhoff, W.; and Kroonenberg, S. B. 1999. Fluvial incision and channel downcutting as a response to late-glacial and Early Holocene climate change: the lower reach of the River Meuse (Maas), The Netherlands. J. Quat. Sci. 14:59–75.

Thornbury, W. D. 1954. Principles of geomorphology. New York, Wiley, 618 p.

Thorne, J. 1994. Constraints on riverine valley incision and the response to sea-level change based on fluid mechanics. *In* Dalrymple, R. W.; Boyd, R.; and Zaitlin, B. A., eds. Incised-valley systems: origin and sedimentary sequences. SEPM Spec. Publ. 51:29–43.

Totterdell, J. M.; Blevin, J. E.; Struckmeyer, H. I. M.; Bradshaw, B. E.; Colwell, J. B.; and Kennard, J. M. 2000. A new sequence framework for the Great Australian Bight: starting with a clean slate. APPEA J., 40:95–117.

van de Graaff, W. J. E.; Crowe, R. W. A.; Bunting, J. A.; and Jackson, M. J. 1977. Relict early Cainozoic drainages in arid Western Australia. Z. Geomorphol. N.F. 21:379–400.

Vandenberghe, J. 1993. Changing fluvial processes under changing periglacial conditions. Z. Geomorphol. N.F. Suppl. 88:17–28.

Waterhouse, J. D.; Commander, D. P.; Prangley, C.; and Backhouse, J. 1994. Newly recognised Eocene sediments in the Beaufort River Palaeochannel. Geol. Surv. West. Aust. Annu. Rev. 1993–1994, p. 82–86.

Williams, G. E., and Gostin, V. A. 2000. Mantle plume uplift in the sedimentary record: origin of kilometredeep canyons within late Neoproterozoic successions, South Australia. J. Geol. Soc. Lond. 157:759–768.

Zachos, J.; Pagini, M.; Thomas, E.; and Billups, K. 2001. Trends, rhythms, and aberrations in global climate 65 Ma to present. Science 292:686–693.

Zaitlin, B. A.; Dalrymple, R. W.; and Boyd, R. 1994. The stratigraphic organisation of incised-valley systems associated with relative sea level change. *In* Dalrymple, R. W.; Boyd, R.; and Zaitlin, B. A., eds. Incised-valley systems: origin and sedimentary sequences. SEPM Spec. Publ. 51:28–60.

Zaitlin, B. A.; Warren, M. J.; Potocki, D. J.; Rosenthal, L.; and Boyd, R. 2002. Depositional styles in a low accommodation foreland basin setting: an example from the Basal Quartz (Lower Cretaceous), southern Alberta. Can. Pet. Geol. Bull. 50:31–72.