

DEPOSITION OF POLAR CARBONATES DURING INTERGLACIAL HIGHSTANDS ON AN EARLY PERMIAN SHELF, TASMANIA

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ABSTRACT: The early Permian Lower Parmeener Supergroup contains two bioclastic limestones that accumulated during the period of prolonged deglaciation of southern Gondwana as the region moved equatorward from 80° S to 70° S. The Darlington and Berriedale limestones formed in neritic environments, in areas where abundant ice-rafted debris testifies to numerous icebergs and glendonites indicating near-freezing seawater. These limestones consist of argillaceous and clean bioclastic floatstone, rudstone, and grainstone that contain a high-abundance, low-diversity heterozoan assemblage of calcareous invertebrates. The components are dominated by large, robust brachiopods, bryozoans, and *Eurydesma* bivalves. Sponge spicules and crinoids are also common, whereas coralline algae and conodonts are conspicuously absent.

Carbonates were deposited on the middle shelf during sea-level highstands, below the iceberg grounding line, where the faunal communities remained undisturbed. In this setting the organisms were outboard of significant terrigenous sediment influx and brackish water, which were trapped on the inner shelf by bathymetry and icebergs. Strong bottom currents also prevented the deposition of fine-grained siliciclastics by continuous winnowing of the pure limestone facies. Upwelling of nutrient-rich water, inferred from the distribution of phosphate, promoted high primary productivity, which fueled this carbonate factory and inhibited dissolution of the biogenic carbonate.

INTRODUCTION

The greater part of carbonate production occurs in photozoan-dominated, tropical, warm-water settings (Bathurst 1975; Tucker and Wright 1990) and much less so in heterozoan-dominated, temperate, cool-water environments (Nelson 1988; James 1997). Part of the reason for this distribution is the increased solubility of CO₂ in colder water, which leads to decreased concentrations of CO₃²⁻, which in turn inhibits CaCO₃ precipitation and sluggish biotic precipitation rates (Mutti and Hallock 2003). These characteristics are even more acute in polar, cold-water settings. Nevertheless, polar limestones do form in modern environments (Taviani et al. 1993; Andruleit et al. 1996), and are inferred to have done so in the past (Beauchamp and Baud 2002). Carbonate strata deposited in these cold environments need to be compared to extant cool-water flora and fauna in order to facilitate their recognition in the rock record. In addition, it is important to study such deposits in terms of the chemical and physical oceanographic mechanisms that allow their precipitation and preservation.

The Lower Parmeener Supergroup (Forsyth et al. 1974) is a 500- to 900-meter-thick marine and terrigenous succession deposited at a high paleolatitude in the Tasmania Basin (Fig. 1) during the late Carboniferous to middle Permian. Marine lithologies consist of both siliciclastic and carbonate rocks that have long been interpreted as cold-water in origin (Banks 1957). Previous work has focused on biostratigraphy (e.g., Clarke and Farmer 1976; Reid 2003) and interpretation of lithofacies and paleoenvironments of specific units (Brill 1956; Clarke and Banks 1975; Brill 1982; Rao and Green 1982). Lithofacies have recently been reviewed, and the stratigraphic nomenclature revised, by

Reid et al. (in press). There is no integrated analysis of these important rocks.

This paper presents the results of a recent detailed sedimentological analysis of core and outcrop across Tasmania. Four phases of deposition are recognized and used to demonstrate the effects of changing sea level, ice cover, and circulation patterns in the Tasmania Basin during early Permian deglaciation, that then provide insight into the mechanisms that enabled carbonate formation.

GEOLOGICAL SETTING

Tectonics

The Pangean supercontinent had largely amalgamated by the late Carboniferous and had gradually migrated northward during the Permian. During this time Tasmania was positioned near the South Pole (Fig. 1), between Antarctica and mainland Australia, and moved northward from approximately 80° S in the Pennsylvanian to 70° S in the Permian (Scotese and Langford 1995; Li and Powell 2001).

The tectonic evolution of eastern Australia and Tasmania during this period was complex, with alternating periods of compression and extension. Prior to the late Carboniferous, during the latter stages of Pangea formation, this region was accretionary as continental fragments collided with the eastern margins and caused regional tectonic shortening (Veevers et al. 1994). In contrast, the late Carboniferous and earliest Permian were marked by overall extension due to thermal subsidence and rifting (Veevers et al. 1994). Compression returned with collision between a volcanic arc and eastern Australia during the Sakmarian but did not

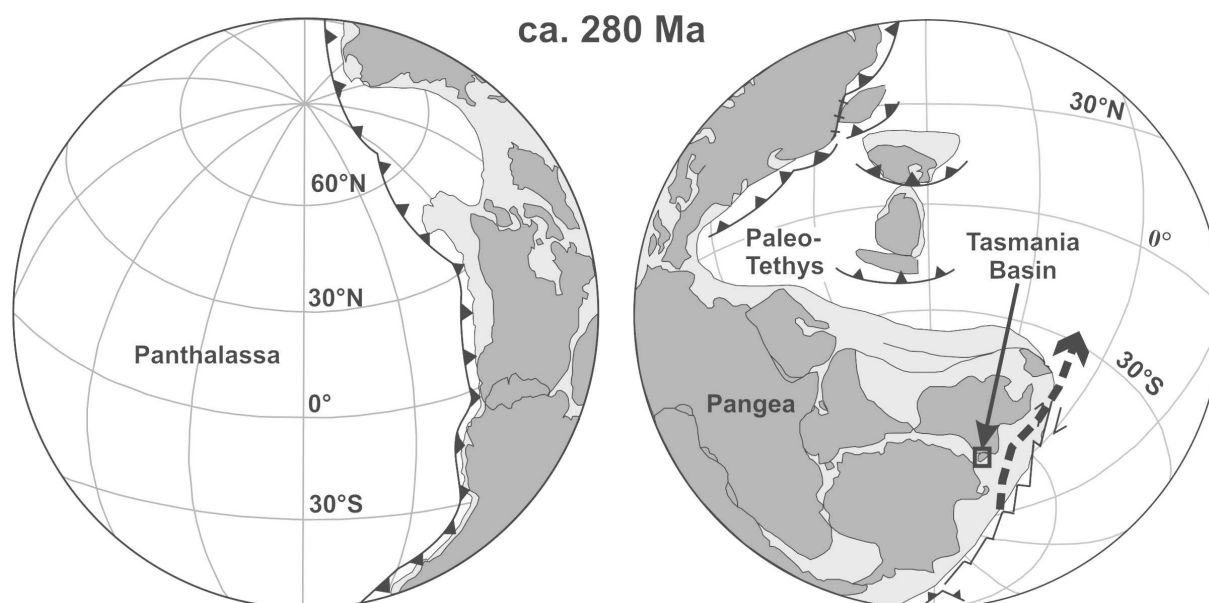


FIG. 1.—Reconstruction of Pangea, from Li and Powell (2001) with the study area shown in rectangle. Present-day continental shelves are shaded dark gray, and the extent of Permian shelves is shown in light gray. The directions of modeled deep ocean current directions are indicated by dashed lines. (Winguth et al. 2002).

affect Tasmania until the Artinskian. Shortening continued throughout the region until the Triassic, when Pangea began to break apart (Veevers et al. 1994).

Climate

Pangean glaciation in the Southern Hemisphere began in the middle Carboniferous (Frakes et al. 1992) and continued with several interglacial–glacial cycles until the mid-Permian (Frakes et al. 1992; Isbell et al. 2003; Jones and Fielding 2004). Recent evidence from Australia and Antarctica suggests that late Paleozoic glaciation was primarily in the form of montane glaciers rather than large continental ice sheets, and that the most extensive glacial peak occurred during the latest Carboniferous–Asselian (Dickins 1996; Isbell et al. 2003).

Southern Pangean climates became progressively warmer as Pangea moved northward throughout the Permian (Ziegler et al. 1997). This climatic amelioration resulted in the growth of *Glossopteris* forests, ferns, and horsetails in Antarctica (Ziegler et al. 1997) and cool-temperate swamps in Australia (Lindsay 1997) by the middle to late Permian. Warming also caused melting of glaciers and concurrent global sea-level rise in the Sakmarian (Dickins 1996). Weather patterns became increasingly seasonal, dominated by strong, intense monsoons along the Tethyan margin (Crowley 1994; Gibbs et al. 2002).

STRATIGRAPHY

Chronostratigraphic boundaries within the Lower Parmeener Super-group are well constrained across Tasmania from both invertebrate and microflora data (Clarke and Banks 1975). This biostratigraphy, however, is difficult to correlate with global timescales because of species provinciality.

Lithostratigraphic names used in the Tasmania Basin (Fig. 2) are numerous, despite overall similar facies relationships. Reid et al. (in press) describe the broad development of facies throughout the basin and propose modifications to the lithostratigraphic nomenclature based on previously established names. This is the format used in this study (Fig. 3).

Marine deposition began in the latest Carboniferous, immediately following glaciation, and persisted into the Sakmarian. Relative sea level dropped, possibly driven by isostatic rebound (Veevers et al. 1994), and terrestrial sediments prograded across the basin in the late Sakmarian. Renewed sea-level rise and marine deposition began in and continued throughout the Artinskian. This inundation was followed by a final major regression in the late Capitanian, presumably due to migration of a peripheral bulge through the area related to an impinging magmatic arc from the east (Veevers et al. 1994).

REGIONAL GEOLOGY

The pre-Permian basement geology of Tasmania can be divided into distinctive western and eastern sectors (Fig. 4). The boundary between the two is traditionally defined by the southern extension of the Tasman Line, represented in Tasmania by a zone of recurring crustal weakness known as the Tamar Fracture System, which also marks the easternmost occurrence of Precambrian rocks (Veevers et al. 1994). New gravity and seismic results, as well as recent mapping, have shown that the Tamar Fracture System more likely denotes a suture between Proterozoic ultramafic rocks in the west and Cambrian mafic and ultramafic oceanic crust in the east (Direen and Crawford 2003). Siliciclastic and carbonate rocks were deposited across this zone during the Cambrian, Ordovician, and Devonian, and all were subsequently intruded by Late Devonian granites (Clarke and Forsyth 1989).

Tasmania was glaciated throughout the Late Pennsylvanian (Clarke and Forsyth 1989; Dickins 1996). Ice sheets are interpreted to have flowed from the west (Fig. 5) and from topographic highs in the northwest, generally moving eastward in topographic lows and converging along the Tamar Fracture System. Glaciers scoured preexisting depressions and led to the deposition of thick marine tillite and glaciolacustrine rhythmites of the Wynyard Formation in fjord-like seaways (Clarke and Forsyth 1989; Hand 1993). This sediment was overlapped by the Asselian Woody Island Formation siltstone, which contains few fossils but abundant glendonite. These fine-grained sedimentary rocks are overlain by the fossiliferous Sakmarian Bundella Formation, which includes the Darlington Limestone (Fig. 3) in the Hobart and Maria Island region. These Sakmarian

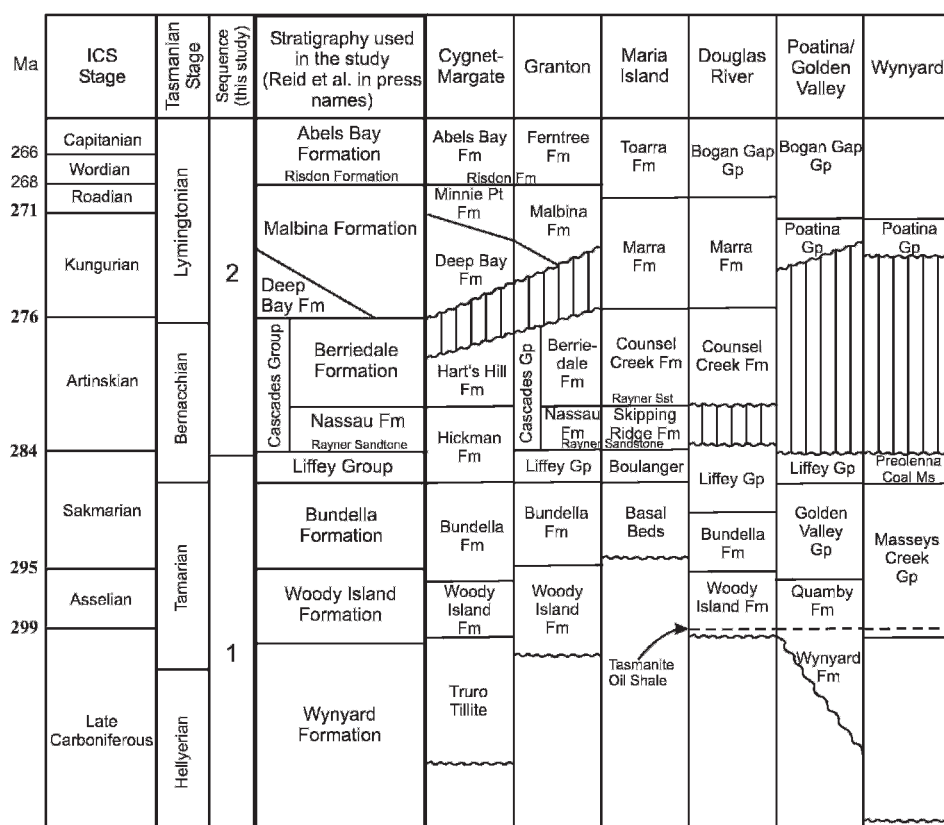


FIG. 2.—Stratigraphic relationships between formations in the Lower Parmeener Supergroup (after Clarke and Forsyth 1989). The locations referred to are shown on Figure 4. Formation names used in this study are provided, corresponding with Reid et al. (in press). Wavy lines and hatched areas represent unconformities.

strata are typified by abundant dropstones and the cold-water *Eurydesma* fauna (Clarke and Banks 1975; Clarke and Forsyth 1989).

These glaciomarine rocks are overlain by late Sakmarian–earliest Artinskian facies of the Liffey Group, which represent a cold-climate, coal-bearing coastal to fluvial siliciclastic succession with minimal evidence of sea ice (Martini and Banks 1989). The pronounced relief produced during glaciation was filled in by Artinskian time, and subsequent sedimentation of the upper marine unit was relatively uniform across the basin. Deposition of the Cascades Group, including the Nassau Siltstone and the Berriedale Limestone, signals renewed marine flooding. Dropstones are abundant in layers within the limestone, and the *Eurydesma* fauna, although present, is much reduced (Clarke and Forsyth 1989). Thin interbedded meta-bentonite layers indicate distant volcanism (Clarke and Forsyth 1989). Sandstone turbidites, variably glauconitic sandstone, and fossiliferous siltstone characterize Kungurian to Capitanian deposition. These strata contain the last occurrences of dropstones and the cold-water *Eurydesma* fauna (Clarke and Forsyth 1989).

This whole succession is overlain by a 600-meter-thick late Permian to Triassic freshwater succession of the Upper Parmeener Supergroup, consisting of estuarine, fluvial, and alluvial sandstones and coal deposits (Clarke and Banks 1975; Clarke and Forsyth 1989). The Parmeener Supergroup was later intruded by Jurassic dolerites during rifting of Pangea (Clarke and Forsyth 1989).

METHODOLOGY

This research is based on sedimentary rocks logged from 18 drill holes across the Tasman Basin, selected for their distribution and lack of alteration (Fig. 4). Drill holes near Granton, Bicheno, Tunbridge, and Eaglehawk Neck were chosen for detailed sampling because of the completeness of the sections they represent. Surface stratigraphic sections were also studied at Mount Nassau, near Hobart, and Maria

Island. Lithofacies were identified in core and confirmed using 213 thin sections.

LITHOFACIES

The Lower Parmeener Supergroup is composed of seven recurring lithofacies associations (attributes summarized in Table 1), exhibiting a general north-to-south, onshore-to-offshore polarity (Fig. 6). For ease of discussion the shelf system is temporally subdivided into: (1) a “segmented inner shelf,” in which antecedent glaciogene topography resulted in an island-and-basin bathymetry on inboard parts of an otherwise homoclinal shelf (Asselian and Sakmarian); and (2) a “continuous inner shelf,” in which previous deposition had largely filled in topography and the surface had a smooth shoreline-to-basin transition (Artinskian).

Diamictite, Rhythmite, and Glendonitic Siltstone Lithofacies Association

The most geographically extensive outcrop of this facies association is found at Wynyard (Fig. 4), where bedded diamictite, consisting of pebble- to block-size clasts in a poorly sorted silt to shale matrix, is overlain by centimeter-scale, locally contorted siltstone and sandstone layers (Fig. 7A). Diamictite beds range from decimeter to decameter scale, and accumulations are reported to range from absent to 450 m thick at Cygnat (Farmer 1985) and more than 550 m at Wynyard (Clarke and Farmer 1976). Beds of massive and laminated siltstone and pebbly mudstone both underlie, and are laterally equivalent to, diamictite (Hand 1993). Southwest of Wynyard the diamictite is structureless or horizontally bedded (Domack et al. 1993). Bedding surfaces are striated, there are local roches moutonnées, and some diamictite has imbricated clasts that are striated in one or more directions (Clarke and Forsyth 1989). Fossils are rare, although brachiopod and bryozoan fragments are

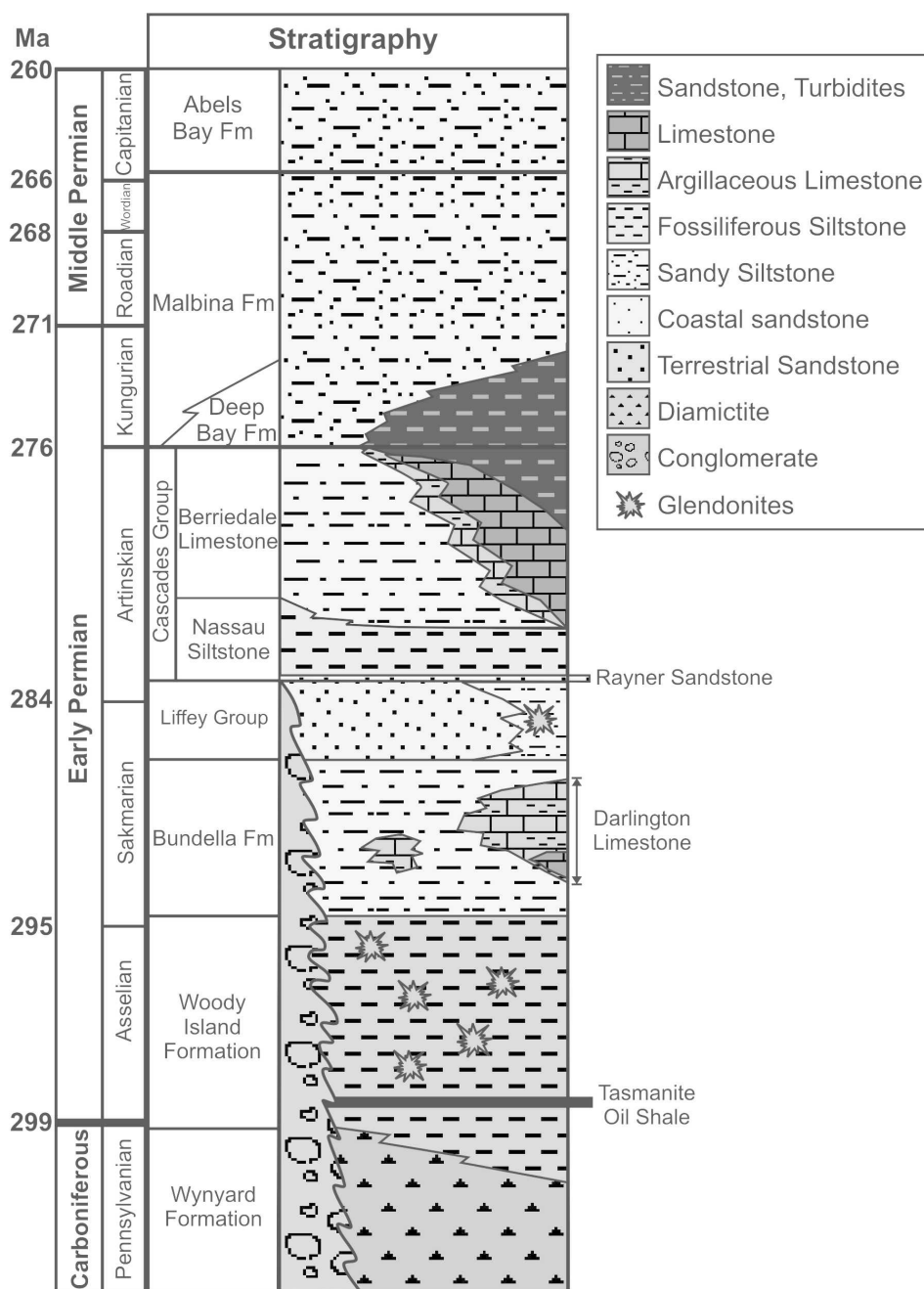


FIG. 3.—Lower Permian Supergroup stratigraphy for Hobart region and central Tasmania showing general stratigraphic distribution of lithofacies (modified from Clarke and Forsyth 1989). Stratigraphic position of Tasmanite oil shale is given despite its absence as a discrete bed in the Hobart region.

present in siltstone beds interbedded with diamictite in drill core from Ross-Quoin, Bicheno, and Eaglehawk Neck (this study) and along the south coast (Hand 1993).

Graded rhythmites are reported from northern Tasmania (Clarke and Forsyth 1989) and from drill core in the south (Hand 1993). The fossil insect *Psychroptilus burrertae*, plants such as *Botrychiopsis plantiana* and *Aphlebia* sp., arthropod tracks of *Tasmanadia twelvetreesi*, and lacustrine spinose acritarchs (Hand 1993) are present in some northern outcrops of the rhythmites. Conversely, shallow marine trace fossils occur in southern exposures of the rhythmites (Hand 1993).

Glendonitic siltstone makes up the Woody Island Formation and part of the Liffey Group. In the Woody Island Formation the fine-grained sediment is pyritic and variably bioturbated. It contains rare dropstones, and abundant glendonite crystals and crystal arrays that can reach several

centimeters in diameter. Rare climbing-ripple cross-stratification, thinly laminated siltstone, and lenticular bedding are reported from the Woody Island Formation in northern Tasmania (Domack et al. 1993). Oil-shale beds, rich in the green alga *Tasmanites*, occur approximately 30 m above the base of the Woody Island Formation and commonly contain dropstones (Clarke and Forsyth 1989; Domack et al. 1993). Individual beds are up to 2 m thick. Glendonitic siltstone, although present, is rare in the Liffey Group (this study), and the glendonite crystals are typically only a few millimeters across (Fig. 7B). In this instance glendonitic siltstone is interbedded with the herringbone cross-stratified sandstone and bioturbated mudstone lithofacies association (see below).

Interpretation.—Diamictite is interpreted as the deposits of glaciomarine debris flows because of their association with marine fossils (Hand

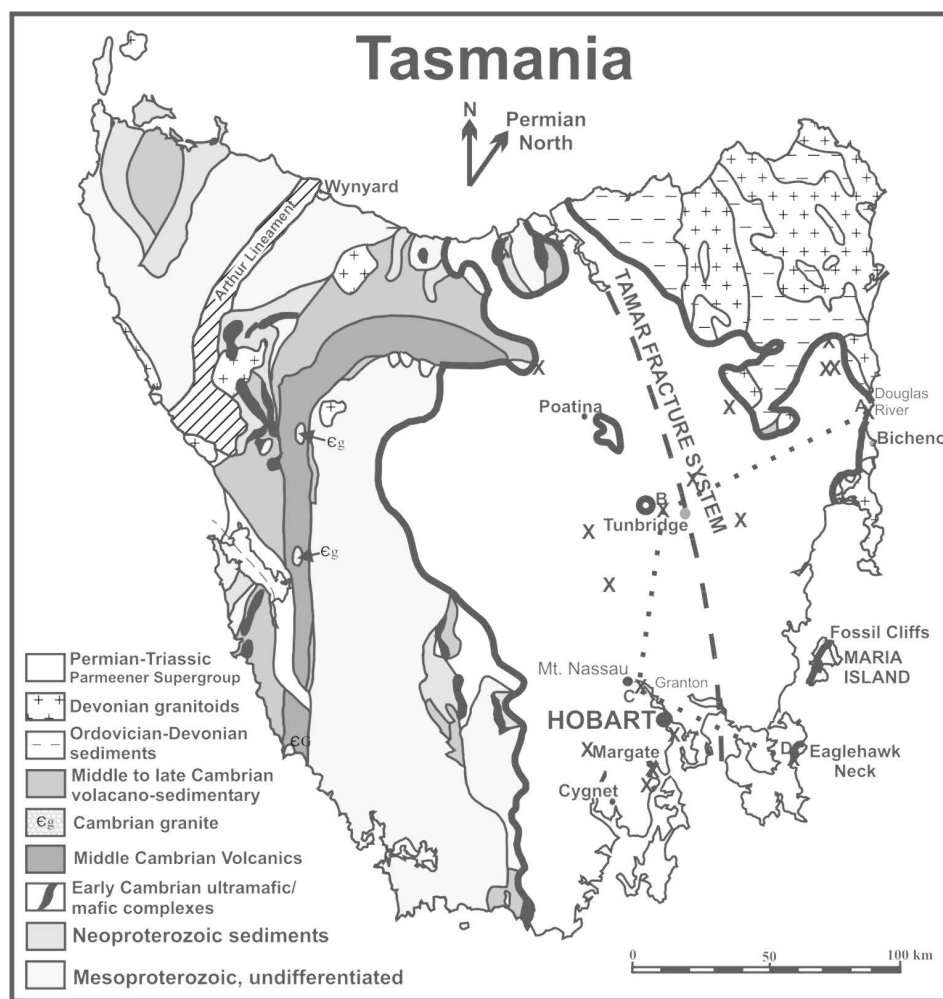


FIG. 4.—Tasmania Geology (from Clarke and Forsyth 1989). Drill holes studied are shown as an “X,” and important cities and outcrop names are shown. Line A–B–C–D shows the position of the cross section in Figure 6 (this study).

1993). Rhythmite beds are inferred to be either glaciolacustrine or glaciomarine on the basis of fossil identification within the beds (Clarke and Forsyth 1989; Hand 1993).

Interpretations of the glendonitic siltstone and Tasmanite oil shale have proven more controversial. Revell et al. (1994), citing the presence of fine-scale lamination and of scouring, proposed that the oil shale was shallow water in origin. Domack et al. (1993) and Domack (1995) disagreed, suggesting that the same features are evidence of distal turbidites and that coarse, pebbly deposits should be associated with nearshore environments in glaciomarine systems, but these are lacking. The geographic distribution of the Tasmanite oil shale supports a deeper-water explanation, although it is proposed here that its accumulation is restricted to isolated sub-basins on the inner shelf. Where the topography was more subdued and sedimentation rates were higher (towards the south), *Tasmanites* is more dispersed and the rocks are composed mostly of glendonitic siltstone.

Overall, Tasmanian glaciomarine diamictite and rhythmites are interpreted to have been deposited in a fjord-like environment, with glaciomarine siltstone first accumulating in the southeast as glacial retreat continued westward (Hand 1993), similar to marine environments in McMurdo Sound today (Bartek and Anderson 1991). McMurdo Sound is characterized by fine-grained sediment, derived from both eolian processes off the top of glaciers and subaqueous rain-out, with minor coarse debris deposited by rafting and turbidity currents originating from the glacier grounding line (Bartek and Anderson 1991).

Cross-Bedded Sandstone, Mudstone, and Coal Lithofacies Association

Martini and Banks (1989) have documented this lithofacies association in detail, and their work provides a contextual basis for the limited sections observed in the drill core in this study, as well as providing detail for associated lithofacies that were not studied.

Meter-scale upwards-fining and -thinning sandstone successions are common, and contain thin pebble layers and cross-bedding at the base that grade up into massive or planar-laminated sandstone beds and cross-laminated siltstone. Some examples of this facies association are dominated by 80-centimeter-thick sandy conglomerate beds. The sandstone and sandy conglomerate lithologies are interbedded with flaser-bedded and organic-rich siltstone (Fig. 7E) containing thin, discontinuous coal seams with a *Glossopteris* flora (Martini and Banks 1989). This lithofacies association is found primarily in the north and northwest parts of the Tasmania Basin, with local occurrences in the northeast.

Interpretation.—These rocks are interpreted as terrestrial debris-flow conglomerate and braided-stream deposits that formed around topographic highs and graded into meandering streams, floodplains, floodplain ponds, and crevasse splays in the valleys (Martini and Banks 1989).

Sandstone and Bioturbated Mudstone Lithofacies Association

This lithofacies association also occurs in the Liffey Group but is characterized by massive, graded, cross-stratified and herringbone cross-

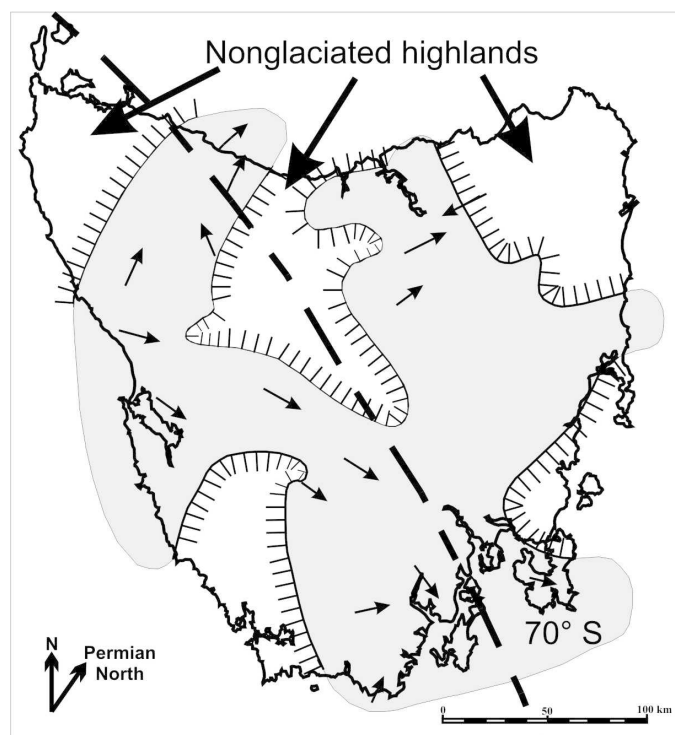


FIG. 5.—Interpreted glaciation and directions of ice movement during the late Carboniferous (after Domack et al. 1993; Hand 1993; Clarke and Forsyth 1989; Reid et al. in press). Glaciated areas are shaded gray, and unglaci-ated highlands are unshaded.

stratified fine- to medium-grained sandstone beds, ranging from 5 cm to 1 m thick. Individual sets of tabular and trough cross-stratification are 1 to 5 cm thick. Sandstone beds are interbedded with bioturbated mudstone beds 2 to 50 cm thick. Laminated mudstone, mudstone drapes, paired mudstone drapes, and flaser bedding are associated with herringbone cross-stratified sandstone that contains some pebbly laminae. These beds exhibit small flame structures and rare ball-and-pillow structures. Interbedded sandy mudstone is heavily bioturbated and commonly contains small *Skolithos* burrows and local *Chondrites* burrows. Plant fragments up to 5 cm long are present on sandstone bedding surfaces (Fig. 7C) and within locally pyritic, carbonaceous mudstone. Centimeter-scale contorted bedding is also present where thin sandstone layers are interbedded with mudstone. Convolutions are 3 to 10 cm high and have the appearance of jagged, sharp-crested anticlines with broad synclines that are separated by several decimeters of offset.

Interpretation.—These facies have been interpreted as peritidal deposits that include intertidal channels and bioturbated mudstone tidal flat facies, and deltaic facies (Martini and Banks 1989). Herringbone cross-stratified sandstone and *Skolithos* burrows indicate deposition of shifting subaqueous dunes along a coastline with tidal influences (Gerdes et al. 2000). This lithofacies association is interbedded with bioturbated mudstone and poorly fossiliferous siltstone from the bioturbated mudstone and fossiliferous siltstone and sandstone lithofacies association (described below). The presence of *Chondrites* burrows indicates high organic content in the sediment. This could be the result of organic input near river mouths (Löwemark et al. 2004), and combined with the abundance of load features may suggest a deltaic environment. The contorted bedding and load structures are similar to sedimentary structures produced by shore-fast ice on tidal flats of the St. Lawrence (Dionne 1998). These features could also be attributed to several other

mechanisms, including, but not limited to, frost action, water escape, and pressure changes from passing storm waves.

Pebbly Sandstone Lithofacies Association

The Rayner Sandstone is a 1-meter-thick, poorly sorted coarse to pebbly sandstone that overlies the herringbone cross-stratified sandstone and bioturbated mudstone lithofacies association and penecontemporaneous marine units from the lower Hickman Formation. It is variably fossiliferous, mainly containing brachiopods, bryozoans, and skeletal fragments, and also contains rounded pebbles ranging from 1 to 5 cm in diameter. This bed is interpreted here to be a shoreface lag deposit, similar to the shoreline described by Mack et al. (2003) from the Permian of New Mexico.

Bioturbated Mudstone and Poorly Fossiliferous Siltstone Lithofacies Association

Bioturbated mudstone and poorly fossiliferous siltstone form the base of the Bundella Formation, are interbedded with sandstone in the Liffey Group and correlative marine Hickman Formation, and dominate the Nassau Siltstone (Fig. 2). These facies are relatively thin across the Tasmania Basin and are most extensive in the central and northern regions. In the north, buff to light brown, massive, sandy siltstone up to 15 m thick, containing pebbles and rare small brachiopods and bryozoans, grades southward into finer-grained facies. These southern facies consist of dark gray, massive, heavily bioturbated siltstone and mudstone ranging from 5 to 25 m thick. The marine biota is dominated by ostracodes and subhorizontal burrows in the central and north part of the basin, and small, thin-shelled brachiopods and diminutive bryozoans in the south.

Interpretation.—This lithofacies association is interpreted here as a nearshore marine deposit. The ostracode-dominated bioturbated mudstone in the north suggests a stressed environment (Curry 1999). The stratigraphic proximity to fluvial facies is interpreted to indicate a freshwater influx sufficient to create nearshore hyposaline conditions that decreased in importance offshore. The bioturbated mudstone grades upward and southward into the poorly fossiliferous siltstone, the latter of which is clearly marine in origin, although the small size of the fossils is again interpreted to reflect environmental stress, probably limited hyposalinity.

Fossiliferous Siltstone and Limestone Lithofacies Association

Medium to dark gray fossiliferous siltstone (Fig. 7D, 8B, C) passes eastward and southward into dark gray limestone with a calcareous silt matrix (Fig. 8A, D), medium gray to buff pure limestone (Fig. 8E, F, G), and spiculitic limestone (Fig. 9).

Fossiliferous Siltstone.—Fossiliferous siltstone occurs in the Bundella Formation, the Berriedale Limestone, and parts of the Nassau Siltstone. It consists of moderately to densely packed fossils in a matrix of massive to heavily bioturbated siltstone to pebbly or sandy siltstone with a floatstone to rudstone texture. The siltstone is relatively homogeneous across 20 m-thick sections of drill core, with variations in fossil abundance and biota on a centimeter- to meter-scale. Fossils are predominantly brachiopods and bryozoans that are larger and more robust than in the poorly fossiliferous siltstone lithofacies, and *Eurydesma* bivalves are also locally present. Brachiopods do appear to be aligned, but bryozoans (Fig. 8B, C) are in all cases horizontally oriented and locally folded. Rare small crinoid ossicles are present in distinct horizons only a few centimeters thick.

Argillaceous Limestone.—Argillaceous limestone, in both the Darlington and the Berriedale, consists of brachiopod-bivalve rudstone and densely packed bryozoan floatstone with a calcisiltite matrix, and forms

TABLE 1.—*Summary of Lithofacies in the Lower Parmeener Supergroup.*

Lithofacies Associations	Description	Fossils	Paleo-environment	Stratigraphic Unit
Diamictite, Rhythmite, and Glendonitic Siltstone	Diamictite: angular to rounded clasts of mixed lithology, striated clasts, 2–50 cm diameter; faceted grains, silt and clay matrix; locally bedded	Rare brachiopods, bryozoans, bivalves	Glaciomarine (Hand 1993; Clarke and Forsyth 1989)	Wynyard Fm
	Rhythmites: sandy siltstone; interbedded medium and coarse-grained sandstone; locally contorted; current-ripple cross-lamination	Rare plant fragments, fossil insects, arthropod tracks, spinose acritarchs, burrows	Glaciomarine and Glaciolacustrine (Hand 1993; Clarke and Forsyth 1989)	Wynyard Fm
	Glendonitic Siltstone: unfossiliferous siltstone; microscopic to 3 cm glendonites; variably pyritic, rare dropstones	<i>Tasmanites</i> oil shale	Glaciomarine (Hand 1993; Clarke and Forsyth 1989)	Woody Island Fm, rare in Liffey Group
Cross-bedded sandstone, conglomerate, mudstone, and coal	Sandy conglomerate, coarse sandstone with rare pebbles; trough cross-bedding	Rare plant fragments	Terrestrial braided and meandering river, floodplain, floodplain ponds (Martini and Banks 1989)	Liffey Group
	Sandstone with fining and thinning upward sequences, cross-bedded base, well-rounded pebbles	Plant fragments common		
	Thin sequences of well-sorted fine sandstone alternating with organic-rich siltstone; ripple cross-stratification, climbing ripples, flaser bedding	Plant fragments, coal		
	Organic-rich mudstone interbedded with coal seams	Plant fragments, coal, root traces		
Sandstone and bioturbated mudstone	Herringbone cross-stratified sandstone, and massive sandstone	<i>Skolithos</i> ichnofacies, ostracodes, rare forams	Tidal channels and flats (Martini and Banks 1989)	Liffey Group
	Bioturbated sandy mudstone interbedded with fine-grained sandstone; flaser bedding present, rare ice deformation features	Few plant fragments		
Pebbly sandstone	Fossiliferous pebbly sandstone to conglomerate, bioturbated	Brachiopods, bryozoan and other skeletal fragments	Shoreface lag deposit	Rayner Sandstone
Bioturbated mudstone and sparsely fossiliferous siltstone	Variably bioturbated mudstone; ice deformation features, rare dropstones	Rare benthic foraminifera and ostracodes in bioturbated mudstone	Inner Shelf	Woody Island Fm, lower Bundella Fm, Hickman Fm, Cascades Gp, Malbina Fm
	Moderate to heavily bioturbated siltstone with sparse fossils and skeletal fragments; dropstones present	1–3 cm thin-shelled brachiopods, diminutive bryozoans, ostracodes, foraminifera in siltstone		
Fossiliferous siltstone and limestone	Moderate to heavily bioturbated siltstone with densely packed fossils and skeletal fragments; dropstones common; rare phosphates	Abundant 3–5 cm thick-shelled brachiopods and small 5–8 cm <i>Eurydesma</i> alternating with bryozoan-rich layers; rare small crinoids	Mid Shelf	Bundella Fm, Darlington Limestone, Cascades Gp, Hickman Fm, Malbina Fm
	Rudstones and dense floatstones with calcareous siltstone matrix; abundant dropstones; phosphates common as nodules and phosphatized bryozoans	Common to abundant 8–10 cm thick-shelled <i>Eurydesma</i> and 3–5 cm thick-shelled brachiopods, common large robust bryozoans, minor crinoids		
	Rudstones, floatstones, and grainstones with cement or minor lime mud matrix; common dropstones; phosphates common to abundant as nodules, phosphatized bryozoans and resedimented grains	Abundant 8–10 cm thick-shelled brachiopods, common bryozoans, crinoids, and plant fragments		
	Spiculitic limestone, phosphatic spiculite; mud or cement matrix; preserved in phosphate nodules	Abundant sponge spicules, small foraminifera; minor to common brachiopod, bryozoan, and crinoid fragments		
Fossiliferous sandstone and siltstone, sandstone turbidites	Interbedded siltstone and fine-grained sandstone, dropstones common but decrease in abundance offshore; discontinuous thin granule conglomerate layers	Common brachiopods and bryozoans	Outer Shelf	Deep Bay Fm

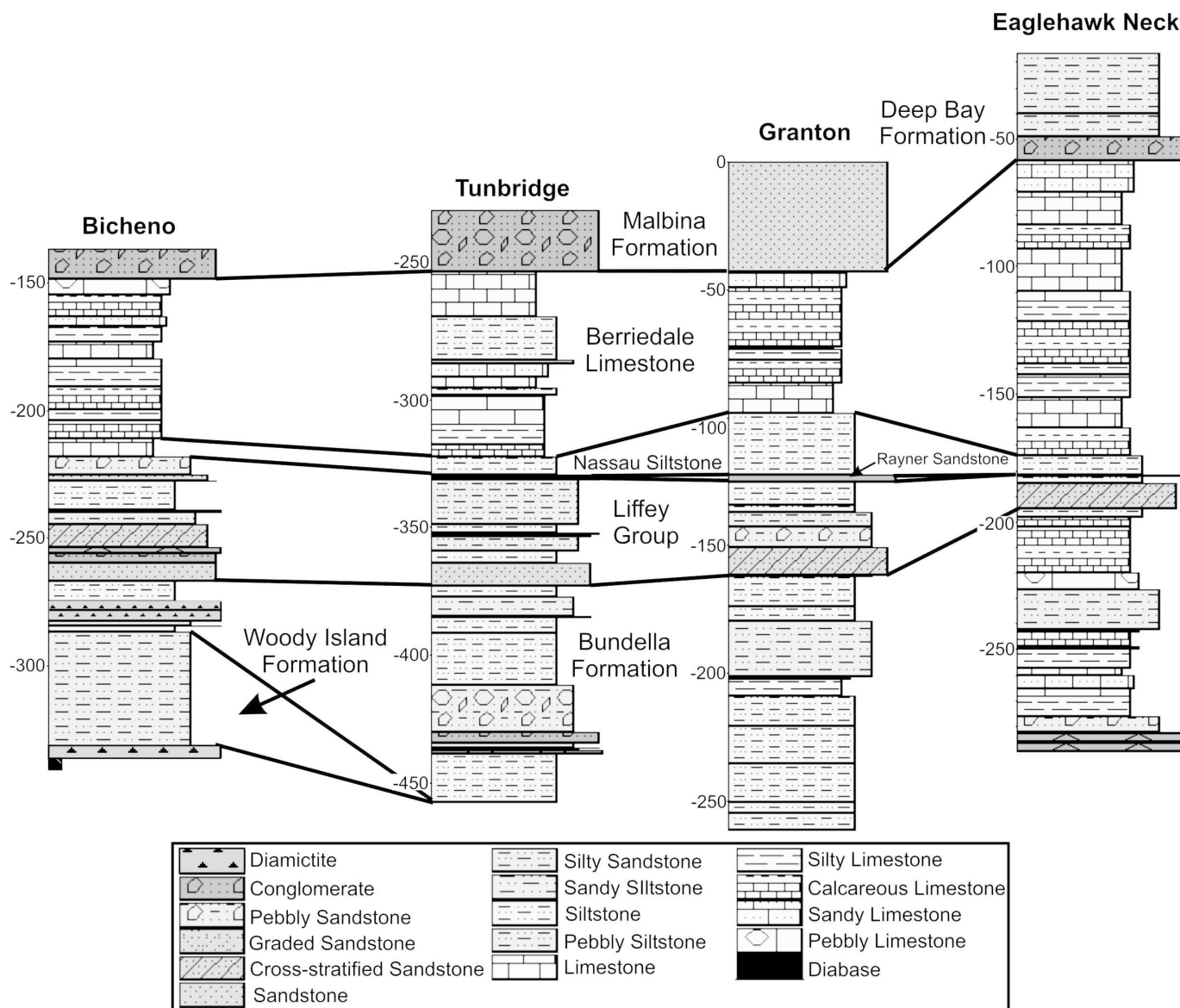


FIG. 6.—Cross section of Lower Parmeener Supergroup along line A–B–C–D (Fig. 3), showing drill holes at A) Bicheno, B) Tunbridge, C) Granton, and D) Eaglehawk Neck.

beds a few centimeters to several meters thick. On Maria Island and a few other isolated localities, outcrops of the Sakmarian Bundella Formation contain *Eurydesma* bivalves, typically bored (Fig. 8D) by barnacles (Reid et al. in press), that form beds up to 3 meters thick and have large-scale cross-bedding (Fig. 8A). A few brachiopods and pectens are also present in such beds. Paleocurrent measurements of cross-bedding and oriented *Eurydesma* shells in beds from Maria Island indicate a unidirectional flow towards the northeast, whereas the shells on the tops of beds show a bidirectional northwest–southeast component approximately perpendicular to the Permian shoreline (Brill 1982; this study). Due to the limited geographic and stratigraphic distribution of paleocurrent data it could not be determined whether these current directions are local or reflect the overall paleocurrent patterns of the Tasmania Basin. Densely packed and well-cemented brachiopod bryozoan quartzose sandy rudstone (Fig. 8A3) forms lenses within the *Eurydesma* beds and becomes progressively more abundant towards the top of the Bundella Formation. Rao (1981) and Brill (1982) report conchoidally fractured quartz in the silt matrix of the argillaceous limestone. Dropstones are numerous and

large, up to 1 meter in diameter, and occur both isolated or in clusters. These dropstones both penetrate underlying beds and crush macrofossils (Fig. 8A1). Phosphate nodules, phosphatized bryozoans, and plant fragments were discovered during this study, but they are not common.

Pure Limestone.—The argillaceous rudstone and floatstone are interbedded with clean limestone in both the Sakmarian Bundella Formation (Darlington Limestone) and the Artinskian Cascades Group (Berriedale Limestone) (Fig. 8E). Pure limestone beds are thicker and more numerous towards the east and southeast. The clean carbonate consists of brachiopod rudstone (Fig. 8D), brachiopod–bryozoan rudstone, and crinoid grainstone, all of which are in beds 10 to 50 cm thick. The matrix, when present, is predominantly productid-spine grainstone, although micrite exists in isolated patches, particularly associated with intragranular pore spaces. Brachiopods, bivalves, bryozoans, and crinoids are typically large and robust compared to those in the argillaceous limestone; dropstones are smaller and less abundant (Fig. 8E). Phosphate is present as dark gray isolated nodules, as replaced bryozoans, and

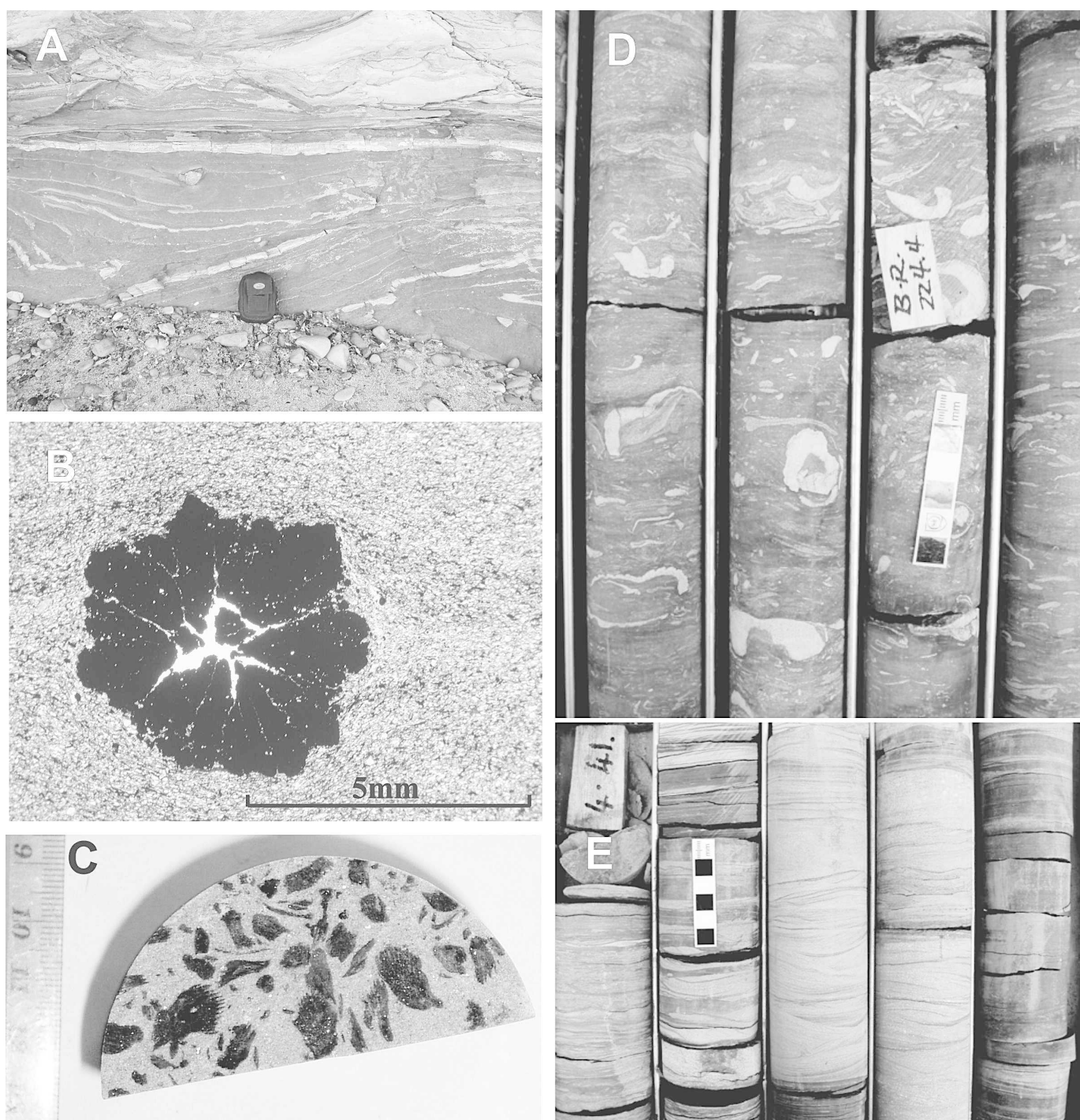
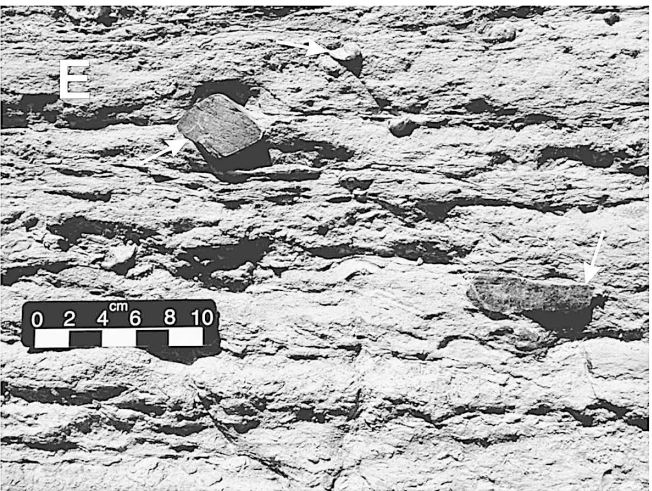
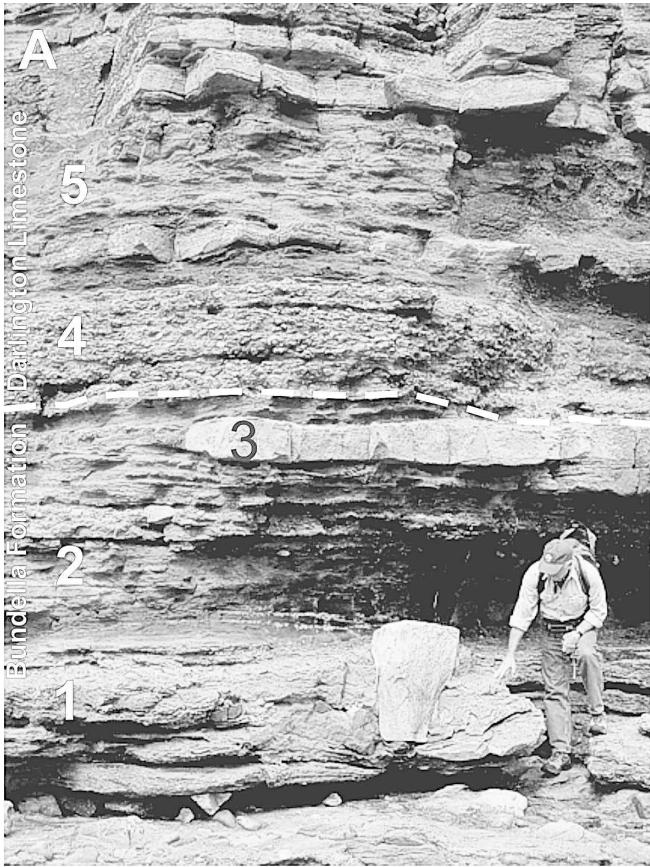


FIG. 7.—**A**) Ice-contact features: contorted sandstone beds in siltstone overlying bedded diamictites at Wynyard; camera case for scale (10 cm high). **B**) Small pyritized glendonite in thin section from the Liffey Group. **C**) Plant fragments in core from the Liffey Group, central Tasmania; scale in centimeters. **D**) Fossiliferous siltstone from the Bundella Formation in drill core near Granton. Scale in centimeters. **E**) Flaser bedding from the Liffey Group in drill core from central Tasmania. Scale in centimeters.

locally as resedimented grains (Fig. 8F, G). Plant fragments are common to abundant in the limestone, especially in crinoidal grainstone.

Spiculitic Limestone.—Spiculitic facies have not previously been described from these rocks. They are found in both the Darlington Limestone and Berriedale Formation, and they are mostly preserved in phosphate nodules where they form phosphatic spiculite to phosphatic-

foraminifera spiculite. Spicules in nodules retain their siliceous composition and grade to spicules replaced by calcite at nodule margins. Stained thin sections (Fig. 9) show evidence of dissolved spicules within pure limestone, replaced with different generations of calcite cement, in quantities large enough to warrant the name spiculitic limestone. Spicules are much more common than has been originally documented and form a significant part of the deep-water biota in the Tasmania Basin.



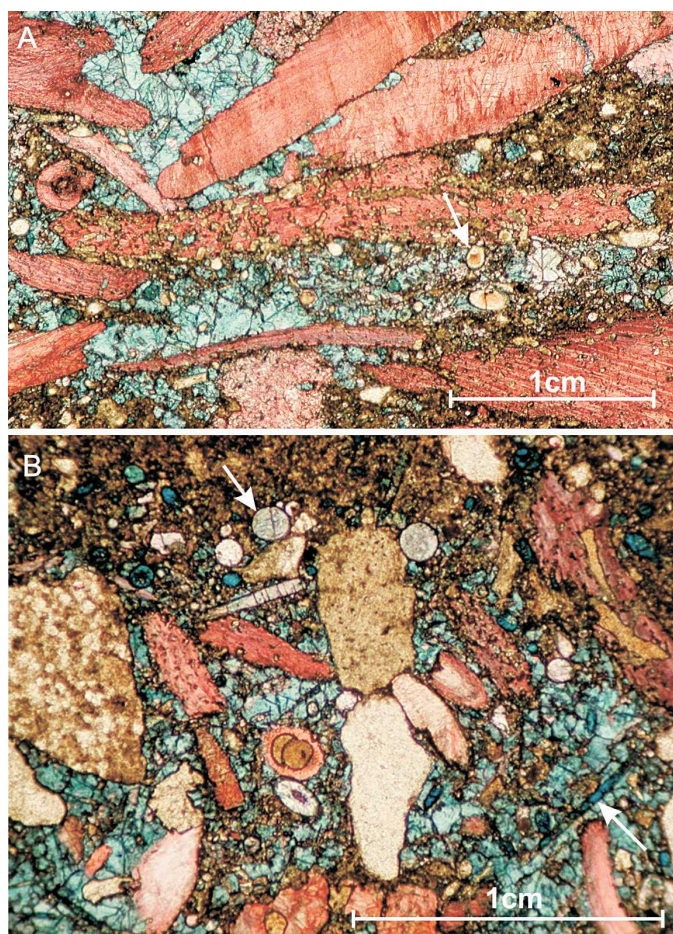


FIG. 9.—Photomicrographs of the spiculitic limestone lithofacies, stained with Alizarin red S and potassium ferricyanide; arrows show examples of spicules. Both sections are from the Berriedale limestone at Eaglehawk Neck. **A**) Siliceous and replaced sponge spicules in phosphatic limestone. **B**) Sponge spicules replaced with late-stage calcite.

Interpretation.—Overall, this progression from poorly fossiliferous siltstone to pure limestone reflects increasingly open marine conditions as facies graded laterally from inner shelf to outer shelf; crinoids indicate normal marine conditions, and the diversity and size of the fauna is greater than in any other lithofacies association. Shell beds with broken and abraded fossils show evidence of episodic reworking, particularly in the argillaceous limestones. *Eurydesma* beds were reworked into shoals on middle-shelf environments (Hobart and Eaglehawk Neck areas) but also formed barrier bars in intertidal areas along headlands (Maria Island) (Brill 1982). Brachiopods and bryozoans become more abundant in interpreted deeper-water facies to the southeast, similar to the distribution seen in modern cool-water environments (James 1997; Lukasik et al. 2000). Spiculitic facies are typically associated with the coldest and

deepest facies both in the Permian Arctic (Beauchamp and Baud 2002) and in modern Antarctic polar carbonate (Taviani et al. 1993).

Fossiliferous Sandstone, Siltstone, and Turbiditic Sandstone Lithofacies Association

This lithofacies association is restricted to the late Artinskian Deep Bay Formation, and the Malbina Formation and equivalent formations (Reid et al. in press). Only limited occurrences were encountered in the drill core used in this study. The Deep Bay Formation is confined to the southwest, and consists of light gray siltstone interbedded with buff, fine-grained sandstone, minor mudstone, and thin discontinuous granule conglomerate laminae. In the northeast, the Deep Bay is coeval with the lower part of the Malbina Formation, as well as the thinner and glauconitic Marra Formation (Clarke and Forsyth 1989). The facies of the upper Malbina Formation, which overlie the Deep Bay Formation in the Hobart region, are fossiliferous pebbly sandstone and siltstone (Farmer 1985).

Interpretation.—The Deep Bay Formation and the lowermost, coeval part of the Malbina Formation are interpreted to represent deposition by sediment gravity flows in the deepest part of the basin (Reid et al. in press). The overlying Malbina Formation is interpreted as a shallow marine, nearshore deposit (Farmer 1985).

GEOGRAPHIC AND STRATIGRAPHIC VARIATION

The Lower Parmeener Supergroup can be divided into four time segments that reflect the progressive lateral and stratigraphic change in lithofacies associations along the shelf profile between the late Carboniferous and the middle Permian.

Phase 1: Late Carboniferous–Asselian

Sediment deposited during this time period reflects regional deglaciation (Fig. 10). Local topographic relief was up to 1 kilometer, and the terrain was rugged (Clarke and Forsyth 1989; Hand 1993), particularly in the Tasmanian interior, and so deposition took place on a segmented inner shelf. As glaciers retreated and sea level rose, deep depressions and U-shaped valleys were filled by glaciomarine diamictite and local glaciolacustrine rhythmite (Wynyard Formation) (Clarke and Forsyth 1989), with coarse conglomeratic facies shed off topographic highs. Clasts were largely derived from northern Tasmanian and Antarctic lithologies (Clarke and Forsyth 1989). Pyritic glaciomarine siltstone (Woody Island Formation) containing large glendonite rosettes and scattered dropstones overlapped these glaciogene rocks across the basin. Tasmanite oil shale accumulated in sub-basins on the inner shelf. Rare shelly biota included *Trigonotreta* brachiopods and *Deltopecten*, *Etheripecten*, and *Eurydesma* bivalves (Clarke and Forsyth 1989).

Phase 2: Sakmarian

The inner shelf remained segmented throughout the Sakmarian (Fig. 11). Bioturbated mudstone and sparsely fossiliferous siltstone floored the inner shelf, grading outboard into mid-shelf fossiliferous

FIG. 8.—**A**) Outcrop at the Fossil Cliffs on Maria Island showing Bundella siltstone (1, 2, 3) and Darlington Limestone (4, 5). Layer 1 is rich in large dropstone boulders crushing underlying layers and bryozoans. Interbedded fossiliferous siltstone and argillaceous limestone (2) are cut by sandy brachiopod bryozoan rudstone channels (3). These beds are interbedded with *Eurydesma* shoals (4) up to 3 meters thick and are overlain by fossiliferous sandy siltstone beds (5). **B**) Large foliaceous trepostome bryozoan from the Bundella Formation on Maria Island. Scale in centimeters. **C**) Large dendroid trepostome bryozoan from the Bundella Formation on Maria Island. **D**) Close-up of *Eurydesma* rudstone in Darlington Limestone showing barnacle borings. **E**) Berriedale Limestone at Rathbone Quarry near Hobart with conspicuous dropstones (arrows). **F**) Phosphate (dark areas) and brachiopods from the Berriedale Limestone at Rathbone Quarry. **G**) Phosphatized trepostome bryozoan in Berriedale Limestone from Tolosa Street Quarry near Hobart.

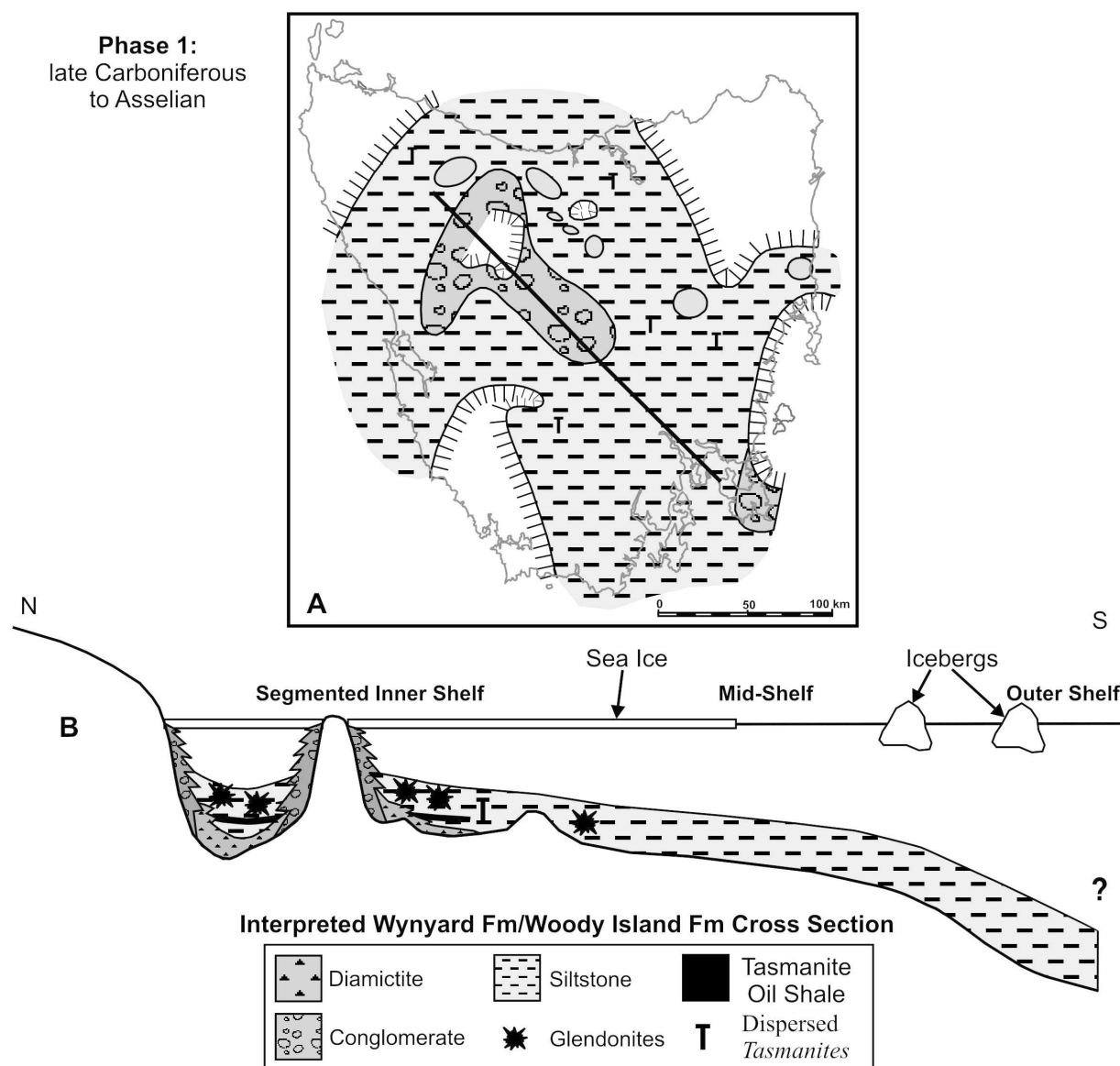


FIG. 10.—**A**) Geographic distribution of Asselian glaciomarine facies (after Reid et al. in press). The line indicates the position of the profile. **B**) Shelf profile of Wynyard and Woody Island Formations during the late Carboniferous to Asselian (this study).

siltstone and argillaceous limestone. Argillaceous limestone was deposited as *Eurydesma* barrier bars that were cut by channels, filled with brachiopod rudstone, in intertidal areas near Maria Island; *Eurydesma* also accumulated as widespread shell lags and shoals (Hobart and Eaglehawk Neck regions). Non-argillaceous limestone was deposited in relative abundance only on the middle shelf near Granton and Eaglehawk Neck and in isolated beds on Maria Island. Phosphate formed locally near Granton in a zone transitional between the argillaceous and the pure limestone. Brachiopods were dominated by *Trigonotreta*, although *Tomioopsis*, *Notospirifer*, *Pseudosyrinx*, *Sulciplica*, *Streptorhynchus*, *Strophosia*, *Pyramus*, *Schzodus*, and *Myonia* were also notably present (Clarke and Forsyth 1989). Bryozoans were abundant, particularly fenestrates and trepostomes (Reid 2003). Plant fragments accumulated in the transitional zone and in non-argillaceous limestone. Dropstones are common throughout all marine facies, but are particularly abundant within argillaceous limestone in the eastern part of the Tasmania Basin.

Some clasts originated in northern Tasmania (Brill 1982), although provenance information is not complete.

Phase 3: Late Sakmarian to Early Artinskian

This was a time of relatively low sea level and widespread terrestrial sedimentation. Alluvial plains prograded across the basin during the late Sakmarian–early Artinskian, finally filling in paleotopography on the segmented inner shelf (Fig. 12). Plant fragments were common in thin coal seams that developed in peat bogs on the plains (Martini and Banks 1989). Herringbone cross-stratified sandstone interbedded with bioturbated mudstone formed in the subtidal environment. Contorted bedding developed at zones of sea-ice contact (cf. Dionne 1998). Glendonite crystals are small and rare, occurring only in marginal marine mudstone from central Tasmania. Inner-shelf facies were predominantly bioturbated mudstone with ostracodes and thin-shelled brachiopods together with sparsely fossiliferous bioturbated sandstone. Brachiopods consisted

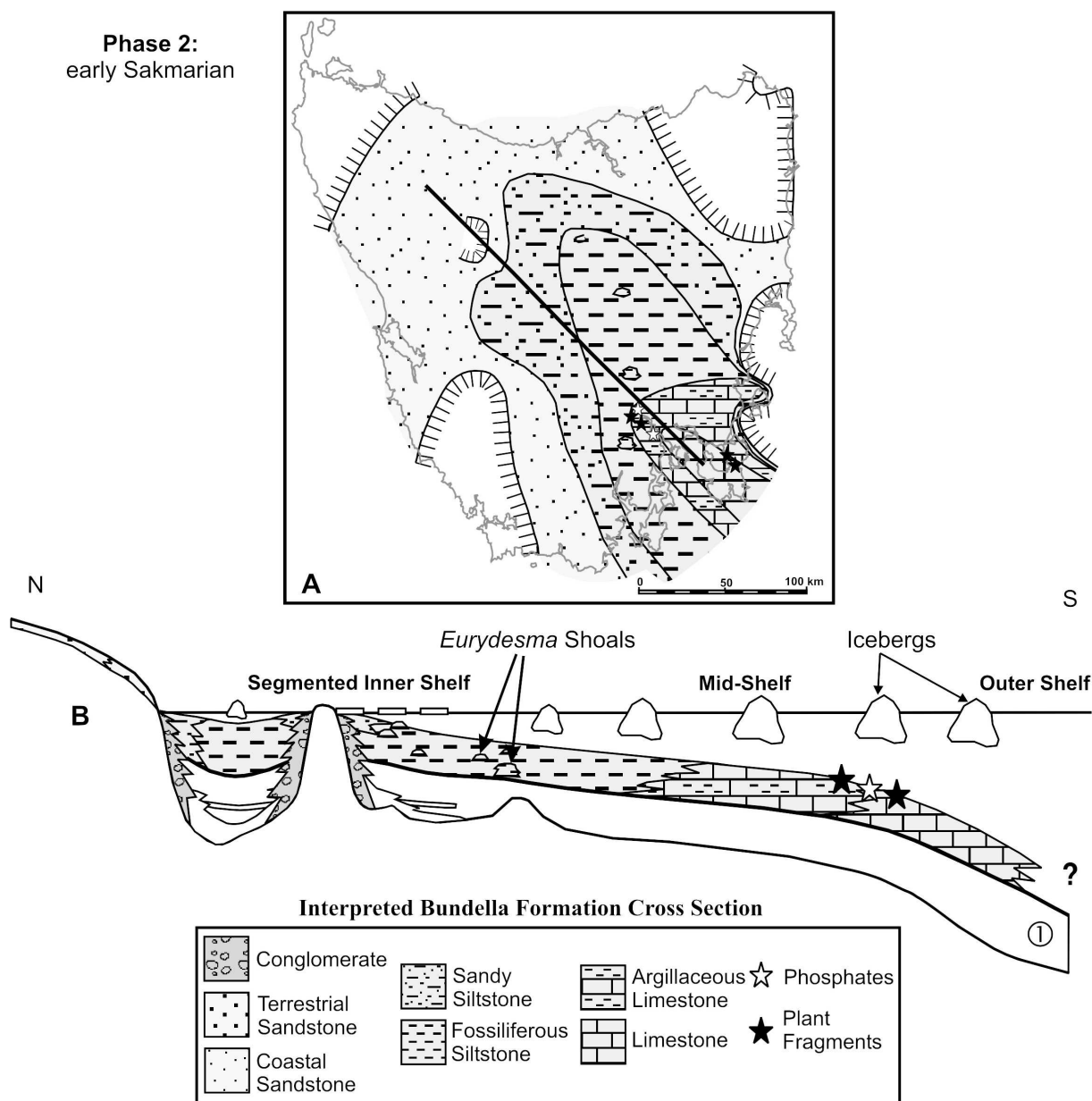


FIG. 11.—**A**) Geographic distribution of Sakmarian Bundella Formation facies (modified from Reid et al. in press). Note the limited distribution of plant fragments and phosphates in the south. The line indicates the position of the profile. **B**) Shelf profile of Bundella Formation and Darlington Limestone during the Sakmarian (this study).

of both spiriferids and productids, predominantly *Tomioopsis* and *Cancrinella* (Clarke and Forsyth 1989). Middle-shelf facies were not developed within the study area during this phase.

Phase 4: Artinskian

The Artinskian was marked by a relative sea-level rise and a return to widespread marine conditions (Fig. 13). The inner shelf was no longer strongly segmented and facies were more uniformly spread across the basin. The nearshore was composed of poorly fossiliferous sand and silt with a few ostracodes and thin-shelled brachiopods. Fossiliferous calcareous silts and argillaceous carbonate graded outboard into pure carbonate that dominated the mid-shelf facies. No *Eurydesma* shoals or barrier bars were present, but winnowed brachiopod and

bryozoan shell beds were common and crinoids became important constituents. During this time the biota was most diverse, dominated by strophalosids and linoproductids with some spiriferids (Clarke and Forsyth 1989). Bryozoans, predominantly fenestrates and trepostomes with rare cryptostomes and cystoporates, were numerous (Reid 2003). Plant fragments and phosphate are abundant in carbonate deposits across the basin. Dropstones are also present but are smaller than in the Sakmarian and generally confined to mid-shelf facies. Fine-sandstone and fossiliferous siltstone turbidites characterized the outer shelf.

SEQUENCE STRATIGRAPHY

Special consideration is required when assessing the sequence stratigraphy of mixed carbonate-siliciclastic systems because it is

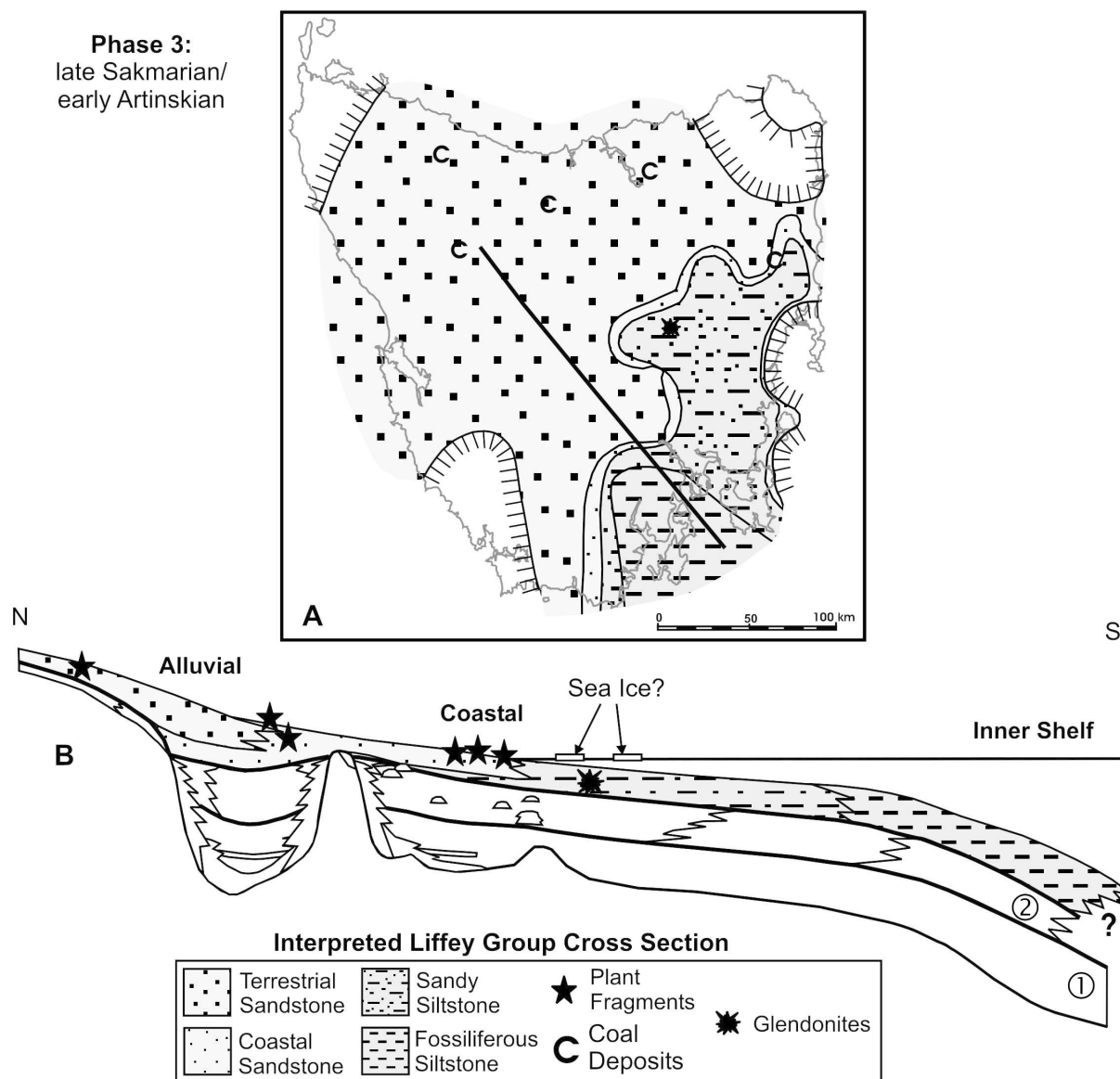


FIG. 12.—**A**) Geographic distribution of terrestrial Liffey Group and marginal marine Hickman Formation (modified from Reid et al. in press). This distribution reflects a marine incursion towards the northeast. Marginal marine facies are typically restricted to the south. The line indicates the position of the profile. **B**) Shelf profile of the Liffey Group and Hickman Formation during the latest Sakmarian (this study).

influenced by the biota as well as by the input of terrigenous sediment (Handford and Loucks 1993; Cunningham and Collins 2002; Caron et al. 2004). In both the cool-water and the cold-water realms there are no ooid shoals and no photozoan assemblages that would produce carbonate buildups. Thus, there is nothing to absorb open ocean wave action, and so deposition is controlled by accommodation space formed by the sea floor and wave base, as in siliciclastic settings. Furthermore, the sequence stratigraphy of glacially influenced basins is additionally complex because of delayed isostatic rebound (Brookfield and Martini 1999; Jin et al. 2002). The Lower Parmeener Supergroup can be resolved into two sequences.

Sequence 1

The base of the first transgressive systems tract is defined by the bedrock glacial erosion surface, which marks the glacial ravinement

surface and sequence boundary (Fig. 14). This surface is overlain by glaciomarine diamictite, rhythmites, and siltstone. In general, maximum flooding surfaces can be identified by the presence of condensed sections reflecting distant sediment sources at the height of transgression (Posamentier and Allen 1999). The maximum flooding surface is difficult to pinpoint in the Woody Island siltstone because there is no single condensed bed. Instead, we suggest that a maximum-flooding zone is represented by the thin series of condensed beds associated with the Tasmanite oil shale within the Woody Island Formation. Above the oil shale the siltstone grades into fossiliferous siltstone and limestone facies, reflecting a gradual deepening in the basin. Highstand limestone deposition typically lags behind maximum flooding (Handford and Loucks 1993) and in the Tasmania Basin does not occur until the Sakmarian. The falling-stage systems tract is interpreted to begin near the end of the Sakmarian at the top of the Bundella Formation, where the lithofacies grade from limestone to fossiliferous siltstone to fossiliferous sandstone and indicate shallowing of the basin. Regression continued

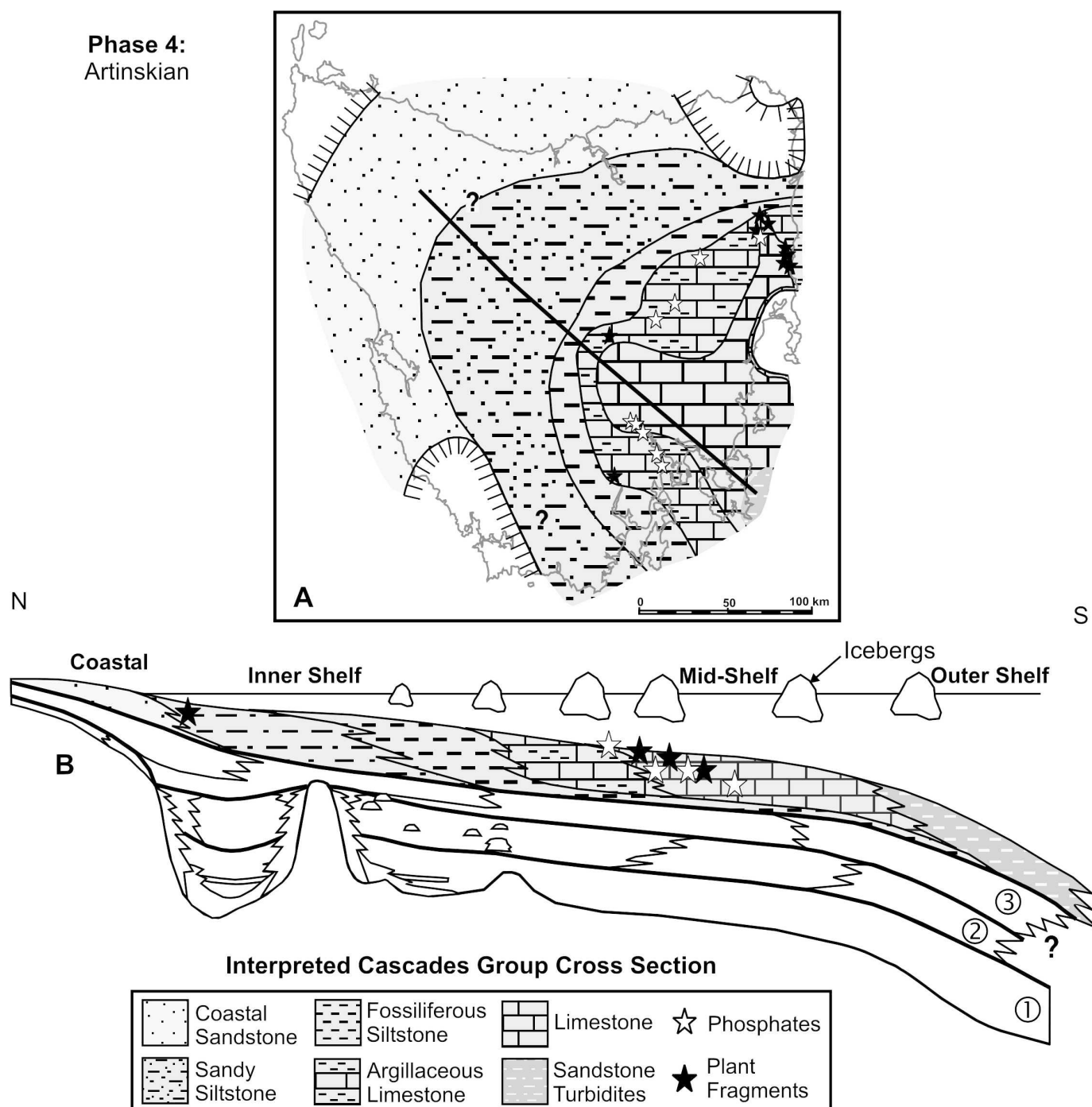


FIG. 13.—**A**) Geographic distribution of Cascades Group (Berriedale Limestone and Nassau Siltstone) facies during the Artinskian (modified from Reid et al. in press). Note phosphates and plant fragments now occur across the mid-shelf. The line indicates the position of the profile. **B**) Shelf profile of Cascades Group and Deep Bay Formation during the late Artinskian (this study).

with the deposition of fluvial facies of the Liffey Group through the end of the Sakmarian and into the earliest Artinskian.

Sequence 2

The Artinskian ravinement surface and sequence boundary is herein defined by an erosional contact between alluvial deposits and marine strata in the north and northeast of Tasmania near Bicheno, Wynyard, and Tunbridge that is coincident with the top of the Liffey Group

(Martini and Banks 1989). The base of the Rayner Sandstone, a thin, fossiliferous pebble conglomerate lag bed, marks the ravinement surface within marine units (cf. Mack et al. 2003). The maximum flooding surface again is not clearly defined in Sequence 2, but it is interpreted to lie at the top of the Nassau Siltstone between the siltstone and limestone beds, similarly to that in Sequence 1. Above this surface there is evidence for condensed beds, such as the phosphate-rich strata in the Berriedale Limestone and the glauconitic sandstone in the Deep Bay Formation, that indicate a stillstand during the highstand systems tract. Above the

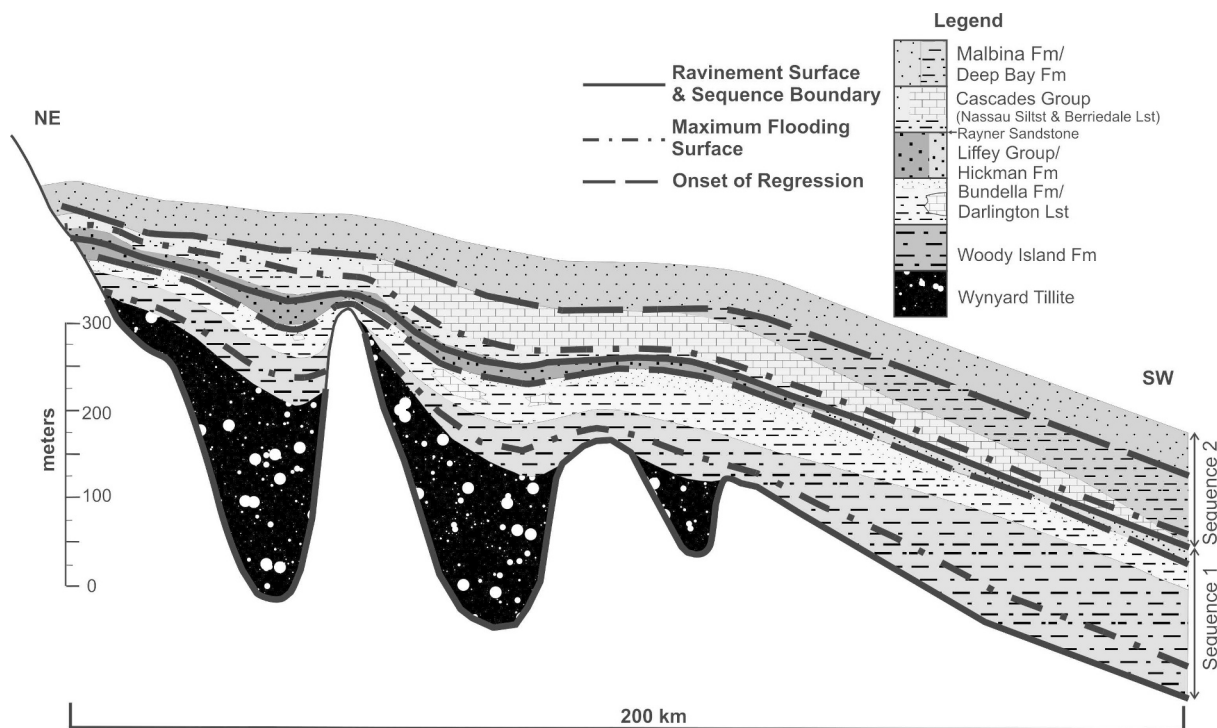


FIG. 14.—Composite facies profile and sequence stratigraphic interpretation of Lower Permian Supergroup shown on an idealized NW to SE cross section. Note that the Rayner sandstone is only 1 m thick and is not portrayed on this diagram.

Deep Bay Formation a final falling-stage systems tract is denoted by deposition of the shallow marine Malbina Formation sandstone overlain by Abels Bay Formation estuarine deposits (Clarke and Forsyth 1989). This sequence culminates in the lowstand terrestrial facies of the Upper Permian Supergroup.

DISCUSSION

Paleoclimate and Paleooceanography

Plant Fragments.—Plant fragments are found throughout the entire deglaciation sedimentary record of the Tasmania Basin, and their relative abundances are an important indicator of climate. Hand (1993) reports rare plant fragments associated with Asselian glaciomarine tillite and glaciolacustrine rhythmites. We also found plant fragments within the Sakmarian Bundella Formation and more commonly in the Artinskian Cascades Group limestone. Thin coal beds are present in the late Sakmarian Liffey Group and thick coal beds in the Upper Permian Supergroup (Clarke and Forsyth 1989). Whereas coal was best developed in interpreted lowstand periods, there is an overall increase in the abundance of plant material upward in the early Permian strata, reflecting the change to a more temperate climate. Clarke and Forsyth (1989) summarize the microflora and note that it became more diverse towards the middle and late Permian. They also identified the macroflora as *Glossopteris*, *Gangamopteris*, and *Noeggerathiopsis*, which would have formed in cold-climate peat bogs similar to the modern Arctic (Hawke et al. 1999) and indicate that the Tasmanian climate became more humid and temperate later in the Permian.

Biotic Abundance and Diversity.—The relative abundance and diversity of microflora, microfauna, and macrofauna also reflect a change in the biologically limiting factors across the shelf through time. Cold temperatures are thought to be the primary influence that caused the

overall low diversity of organisms in the Tasmania Basin during the early Permian (Clarke and Farmer 1976). Biotic abundance and diversity are at their lowest in the Woody Island Formation and the Liffey Group (and coeval marine correlates of the Hickman Formation). Whereas periodic sea-ice cover could have contributed to this situation during the Asselian, fresh-water influence is interpreted to have been important during the late Sakmarian. Phosphate is absent from both the Woody Island Formation and Liffey Group, implying that upwelling was minor and, therefore, the basin overall had normal marine water with relatively low trophic resources.

Water Temperature.—The Tasmanian ocean is considered to have been cold during the Permian (Ziegler et al. 1997). Oxygen isotope studies of bivalve and brachiopod shells by Rao (1981, 1983, 1988) and Rao and Green (1982) from both the Darlington Limestone and the Berriedale Formation limestone deposits have been used to estimate paleotemperature. Data show a strong overlap between the two limestone units along a meteoric alteration trend, with estimated ocean water temperatures comparable to modern Antarctic water near 0°C (cf. Taviani et al. 1993).

These cold marine temperatures are confirmed by the occurrence of glendonite, pseudomorphs of ikaite (Suess et al. 1982), within the siltstone of the Woody Island Formation and the Liffey Group. Ikaite, $\text{CaCO}_3 \cdot 6\text{H}_2\text{O}$, is precipitated in water temperatures near 0°C in marine and nonmarine environments in quiet-water, organic-rich sediment (Suess et al. 1982; Jansen et al. 1987), in mixing zones (Whiticar and Suess 1998; DeLurio and Frakes 1999; Buchardt et al. 2001; Omelon et al. 2001), and at methane seeps (Greinert and Derkachev 2004). The presence of phosphate and alkaline water increases the stability of ikaite (DeLurio and Frakes 1999; Bischoff et al. 1993) and sulfate reduction (Greinert and Derkachev 2004). Quiet-water, organic-rich sediment is envisioned for the Tasmanian glendonites, with enhanced precipitation due to increased alkalinity from decomposing organic matter.

While both of these temperature indicators suggest that the water remained very cold throughout deposition of the Lower Parmeener Supergroup, lack of glendonites in the limestone may indicate slightly warmer temperatures during their deposition or simply be a result of higher energy. Ikaite forming in modern Arctic springs shows seasonal alteration zoning and can decompose into a “crystal mush” at temperatures as cold as 0°C (Omelon et al. 2001). The high energy present during deposition of the Darlington and Berriedale limestones would have physically destroyed any evidence of these delicate crystals.

Dropstones and Ice Cover.—Dropstones are common in the Lower Parmeener Supergroup, particularly associated with the limestone units. They are, however, rare in Asselian glaciomarine siltstones. Temporally, dropstones are most abundant in Sakmarian and Artinskian deposits, particularly in mid-shelf silty limestone.

These outsized clasts have always been interpreted as dropstones derived from icebergs, but there are several other mechanisms that must be ruled out, such as biological rafting and sea ice (Bennett et al. 1996). While tree-borne clasts up to 3 m have been reported (Bennett et al. 1996), outsized clasts transported by plants are typically 3 to 100 mm in size, even in the Amazon, where trees are quite large (Doublet and Garcia 2004). Glossopterids were the size of small shrubs or small trees (Taylor and Taylor 1993), so it is doubtful that they were capable of transporting the large dropstones found in the Lower Parmeener Supergroup. Transportation by sea ice is more difficult to distinguish from iceberg rafting, although the large size, striations, and abundance of clasts within the Lower Parmeener Supergroup are more consistent with icebergs (cf. Bennett et al. 1996; Dowdeswell et al. 1998).

The Asselian should have been a time of abundant icebergs and possibly sea ice because it followed a period of glaciation, yet dropstones are rare, implying that there must have been a mechanism to minimize the occurrence of icebergs in the Tasmania Basin. Sikussaks, thick accumulations of sea ice, described from modern Greenland fjords are effective at trapping icebergs (Syvitski et al. 1996), and such features may have been present in Tasmania. Sea ice serves to slow the deposition of coarse iceberg debris and promotes the deposition of the silt and sand fraction. Development of sikussaks is generally found along grounding lines and in embayments (Syvitski et al. 1996), so segmentation of the Tasmanian inner shelf may have been important in this regard.

The occurrence of *Tasmanites*-rich beds in the upper Carboniferous lower Woody Island Formation is also interpreted to indicate the presence of sea-ice cover. Revill et al. (1994) suggest that *Tasmanites*, a green alga, inhabited sea ice, similarly to living relatives. The *Tasmanites* beds, however, are also associated with dropstones (Domack et al. 1993). The inferred presence of icebergs suggests melting of sea ice in the Tasmania Basin in order to allow the icebergs to enter the basin. The melting of the sea ice would have released the *Tasmanites* and allowed it to accumulate on the seafloor. For this interpretation to be plausible the Tasmania Basin had to be a stratified water mass so that the algae accumulations could be preserved (Palliani et al. 2002). Seawater stratification is supported by the abundance of pyrite in the siltstone, indicating reducing bottom waters, and the generally poor benthic fauna. Development of a fresh-water lens similar to situations in some Holocene fjords (Tamburni et al. 2002; Dallimore et al. 2005), combined with periodic sea ice and the segmentation of the inner shelf, could have inhibited circulation (Ogorodov et al. 2005) and caused stratification of the water column.

The Sakmarian Bundella Formation environment is interpreted to have been similar to the modern Antarctic and Arctic, dominated by icebergs and possibly seasonal sea-ice cover (cf. Syvitski et al. 1996; Stemmerik 2000). The sikkusak would have disappeared by this time, allowing icebergs to move freely through the basin, although segmentation of the inner shelf probably prevented icebergs from entering some sub-basins.

Dropstones are generally less numerous, smaller, and more widespread across the ancient shelf in the Artinskian (Cascades Group and Deep Bay Formation), and there is no direct evidence of a nearby glacial source. This situation is interpreted to indicate periodic movement of icebergs along the coast and into the basin, similar to conditions in the Holocene in the modern North Atlantic (Dowdeswell et al. 1998), where interglacial periods experience only sparse, seasonal iceberg and sea-ice rafting derived from distant glaciers.

Salinity.—Salinity stress may have been important during deposition of the Lower Parmeener Supergroup. Sea ice is interpreted to have covered much of the Tasmania Basin during the Asselian. Such freezing of ocean water would have increased water salinity on the inner shelf, possibly promoting glendonite formation (Ogorodov et al. 2005; Papadimitriou et al. 2003), although this was probably important in the Tasmania Basin only during the Asselian.

During the sea-ice-free periods of the Sakmarian and Artinskian (Phases 2 and 4), inner-shelf sediment would have been influenced by runoff and fluvial influx, as well as melting of icebergs. This freshwater is interpreted to have had the combined effect of creating a lens of somewhat brackish water and a relatively high suspended-sediment load, such as in the Kangerdlugssuaq Fjord in Greenland (Syvitski et al. 1996). These two factors likely explain the abundance of ostracodes and the absence of other filter-feeding organisms. There is nevertheless an overall increase in the faunal diversity during the early Permian as Pangea moved north and the environment became more temperate (Lindsay 1997; Ziegler et al. 1997).

Phosphate.—Phosphate is scarce in Sequence 1 on the Sakmarian segmented shelf, occurring only in silty carbonate of the Darlington Limestone and sandy siltstone of the upper Bundella Formation, and even in these units it is localized to the southwestern part of the basin near Hobart (Fig. 12). This distribution is interpreted to be the result of active, but localized, upwelling.

In contrast, the Artinskian strata have a greater amount of phosphate across the interpreted middle shelf (Fig. 13). At that time the shelf was non-segmented, and thus there was little impediment to circulation. *Eurydesma* bivalves were present but their shells were not reworked into large bars as they were at Maria Island in the Sakmarian. This is possibly because of decreased bivalve abundance or increased wave activity across the shelf due to reduced bathymetric complexity. Crinoids are also much more prevalent in the Artinskian limestones, indicating normal marine conditions and, therefore, a well-mixed water column. Overall, upwelling is interpreted to have been prevalent, with open circulation carrying nutrients inboard across the middle shelf.

Circulation.—Middle to late Permian ocean-circulation models hypothesize a strong, deep ocean current that moved northward along the east coast of Pangea (Fig. 1), with upwelling thought to occur at approximately 70° latitude (Winguth et al. 2002). This model is substantiated by the presence of phosphate in the Tasmania Basin (Rogala et al. 2005) and the distribution of brachiopod provinces during the early and middle Permian (Shi and Archbold 1996). Cold-water brachiopods of the lower Permian Indoralian Province extend from Tasmania through New Zealand, the Shan-Thai terrane, and India, and persist as a cold-water to cool-water brachiopod assemblage until the end of the Permian, indicating the existence of cold currents throughout. Archbold (1997, 2000) postulates that these cold-water currents actually intensified and reached progressively higher latitudes between the Sakmarian and Artinskian, and this interpretation explains the persistence of cold-water marine conditions during terrestrial warming trends recorded elsewhere in Australia (Jones et al. 2006).

Circulation within the Tasmania Basin supports these hypothesized global patterns. During the Asselian (Phase 1), sea ice is thought to have covered most of the Tasmania Basin, although there were relatively rare periods of open water. Prevalence of silt-size sediment, combined with the presence of glendonite and pyrite, as well as restricted biota, indicate a stratified water body and low-energy conditions. Icebergs were prevented from moving across the shelf, as inferred from the scarcity of dropstones. The most likely explanation for this within a glaciomarine setting would be to have had widespread sea ice.

Open-water conditions prevailed during the Sakmarian (Phase 2), although the circulation of nutrient-rich currents was not able to penetrate the entire basin. Phosphate deposits exist near Granton and Hobart, but nowhere in the east. Dropstones were also more abundant in the west, implying that islands blocked or redirected upwelling currents. *Eurydesma* shoals, while irregularly distributed, could also have influenced water movement.

Sea level dropped at the end of the Sakmarian (Phase 3), preventing deep-water currents from permeating onto the shelf. Instead, the interaction of freshwater influx from rivers, tidal currents, and wave activity governed oceanographic conditions on the shallow shelf.

Sea level rose once again during the early Artinskian (Phase 4). Irregular basin bathymetry was filled in during Phase 3 and upwelling currents were free to flow across the Tasmania Basin, as indicated by the widespread distribution of phosphate. Dropstones were also geographically more widespread, although they were smaller. Because the basin did not have as many barriers within it, icebergs were free to float across the shelf wherever it was deep enough. Furthermore, these icebergs may have originated from outside the Tasmania Basin, traveling northward from Antarctica.

Mechanisms of CaCO₃ Deposition in Cold Water

Deposition of carbonate in cold water requires several criteria that enhance precipitation and inhibit dissolution. Cold, polar seawater is undersaturated with respect to CaCO₃ in comparison to warm, tropical seawater, and thus organisms living in cold water are required to expend more energy to extract ions to build and maintain shells (Mutti and Hallock 2003). Furthermore, fresh meltwater in glaciomarine settings enhances undersaturation of CaCO₃, and the corresponding deposition of terrigenous material would dilute any carbonate that did precipitate. In the case of the Lower Parmeener Supergroup carbonate, upwelling of nutrient-rich water provided the energy resources for building shells by increasing primary productivity. Episodic storm and wave energy would have reduced the sedimentation rate in mid-shelf regions.

Besides the difficulty of initially precipitating carbonate material, shells must also be able to withstand dissolution in cold, corrosive water. The main factors controlling dissolution of biogenic carbonate, other than water chemistry, appear to be total surface area, grain size, and microstructure (Walter and Morse 1984; Harper 2000), such that an allochem with a smaller surface area, larger size, and simpler microstructure with large crystals is more resistant to dissolution. This is the case for large, robust brachiopods and *Eurydesma* bivalves that are the best preserved specimens in the Tasmanian limestone.

The presence of phosphate derived from upwelling water may also have been instrumental in preserving the Sakmarian Darlington and Artinskian Berriedale limestones. Even minor amounts of phosphate can inhibit the dissolution of calcite (Berner and Morse 1974; Morse 1974), and it was clearly abundant during deposition of the Tasmanian carbonate.

Polar Limestone

Tasmanian polar carbonate was deposited within mid-shelf facies during highstand conditions in the Sakmarian and Artinskian. Bryozoans, brachiopods, and crinoids dominated the non-argillaceous grainstone and rudstone deposits, indicating bottom currents that were sufficient to

sustain a filter-feeding community, rework the sediment, and prevent the accumulation of mud. This setting is similar to the environments that support the formation of Holocene and modern polar carbonate in the Arctic (Andruleit et al. 1996) and Antarctic (Taviani et al. 1993).

Sponge spicules are preserved in phosphatized nodules in the Berriedale Limestone and the Malbina Formation and as molds in spiculitic limestone. Siliceous sponge body fossils and spicules are frequently found in modern cold-water environments (Taviani et al. 1993) and are used as indicators for such environments in the rock record (Beauchamp and Baud 2002). Their relative scarcity in the Tasmanian deposits can probably be explained by dissolution during diagenesis and reprecipitation as chert in the Darlington and Berriedale limestones.

Open water and icebergs, as indicated by dropstones, prevailed during periods of carbonate production. This oceanographic setting would have supported moderate to high primary productivity, resulting from input of nutrients from debris shed off icebergs. Several studies have recently documented the devastation iceberg grounding has on a variety of benthic communities (Gutt and Starman 2001; Gerdes et al. 2003), and such grounding could have prevented the establishment of a carbonate factory on the inner shelf. Filter-feeding assemblages, similar to those recorded in the Tasmanian limestone, are particularly slow in re-inhabiting such sites (Gutt and Starman 2001; Gerdes et al. 2003). Carbonate production, therefore, was concentrated on the middle shelf, probably because the seafloor lay below iceberg grounding depths.

Carbonate production was likely promoted by the upwelling of nutrient-rich water onto the shelf. Upwelling was limited during the Sakmarian, as reflected by the restricted distribution of phosphate near Hobart and Eaglehawk Neck, and the comparatively thin nature of the Darlington Limestone away from the localized intertidal barrier-bar deposits developed at Maria Island. Artinskian upwelling and phosphate production was more extensive, mirrored by the production of abundant pure limestone across the shelf (Berriedale Limestone, in the Cascades Group). This relationship between upwelling and the formation of polar carbonate appears to be important not only in the Permian of Tasmania but also in more recent examples from the Antarctic (Taviani et al. 1993) and the Arctic (Andruleit et al. 1996).

Comparison with Cool-Water Carbonate

Cool-water carbonate is characterized by a heterozoan biotic assemblage, which precludes inorganic precipitates, green algae, and invertebrates with photosymbionts (Nelson 1988; James 1997). Molluscs, benthic foraminifera, coralline algae, and sponges typify inner-shelf facies, whereas outer-shelf facies contain abundant bryozoans, brachiopods, planktic and benthic foraminifera, echinoids, barnacles, and sponges (James 1997). Permian cool-water carbonate faunas would also contain conodonts but no calcareous plankton, inasmuch as they had not evolved yet, and barnacles would not have been common (Anderson 1994).

The Tasmanian cold-water assemblage is superficially similar to that of cool-water carbonate, but with a few exceptions. In the Tasmania Basin there are no conodonts or coralline algae in any facies. Recent recognition of sponge spicules preserved in phosphate nodules, and as molds in limestone, indicates that siliceous sponges were an important part of the Tasmanian polar biota. The *Eurydesma* fauna is a unique addition to the Permian polar assemblage, which is specifically a Gondwanan fauna associated with cold water (Runnegar 1979). Besides distinctive faunal components, the Tasmanian cold-water carbonate is associated with dropstones and glendonite indicating the presence of ice and cold water.

CONCLUSIONS

Lower Parmeener Supergroup strata are interpreted to have been deposited on a polar cold-water shelf as two depositional sequences.

Partitioning of the basin by antecedent topography during the late Carboniferous to Asselian, and very cold seawater temperatures, are interpreted to have promoted the accumulation of shore-fast sea ice, possibly resulting in increased salinity and reduced mixing of the water column. Post-Asselian inner-shelf environments are characterized by bioturbated mudstone, siltstone, and sandstone with scarce shelly fossils consisting of ostracodes, benthic foraminifera, small thin-shelled brachiopods, and diminutive bryozoans. The paucity of organisms at the time is thought to have been a result of hyposalinity from fluvially derived fresh water. Middle-shelf facies grade outboard from fossiliferous siltstone to limestone to spiculitic limestone and are dominated by brachiopods, bryozoans, crinoids, and lesser bivalves. Plant fragments, phosphate, and dropstones are concentrated in mid-shelf facies. Outer-shelf facies are preserved only in the Artinskian strata of the Deep Bay Formation, and are characterized by siliciclastic turbidites and fossiliferous siltstone and sandstone.

The early Permian Tasmanian climate was cold during postglacial deposition but became progressively more humid and temperate towards the late Permian. Water temperatures remained cold throughout deposition of the Lower Permian Supergroup, although surficial ice conditions changed. Two paleoceanographic states are interpreted to have existed during the early Permian: (1) sea-ice-covered, and (2) open-water with icebergs. Siltstone deposition beneath sea ice characterized the Asselian, whereas mixed siliciclastic-carbonate deposition occurred during the Sakmarian and Artinskian, when open water and iceberg conditions prevailed. Possible hypersaline conditions developed beneath sea ice during the Asselian, although most of the early Permian experienced a gradient of hyposaline to normal marine conditions offshore as a result of fluvial freshwater influx. Upwelling, as demonstrated by phosphate deposition, was important on mid-shelf facies during periods of open water during the Sakmarian and Artinskian. Upwelling was most widespread through the basin in the Artinskian, when islands on the shelf did not obstruct circulation.

Middle-shelf Tasmanian limestone was deposited under open-water conditions with icebergs during highstands. Carbonate production was enhanced below the iceberg grounding line by upwelling of nutrient-rich water and by strong currents.

Tasmanian polar carbonate is characterized by a heterozoan assemblage similar to that of cool-water carbonate. In addition, this polar carbonate contains the Gondwana *Eurydesma* fauna, sponge spicules, glendonites, and dropstones. Conodonts and coralline algae are absent.

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