THE UNIVERSITY OF CHICAGO

LATE CRETACEOUS TO PLEISTOCENE CLIMATES: NATURE OF THE TRANSITION FROM A 'HOT-HOUSE' TO AN 'ICE-HOUSE' WORLD

VOLUME ONE

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CHAPTER II

THE GEOLOGICAL EVIDENCE FOR TRIASSIC TO PLEISTOCENE GLACIATIONS: IMPLICATIONS FOR EUSTACY[1]

"If nothing else works, then a total pigheaded unwillingness to look facts in the face will see us through"

"Private Plane"
"Blackadder goes forth. Part 2." BBC, 1990

ABSTRACT

Evidence for Mesozoic and early Cenozoic glaciations is ambiguous at best. Most of the purported evidence including faunal changes, glendonites and stable isotope paleothermometry imply cold conditions but do not explicitly require ice-sheets. In the absence of established indicators of glaciation, such as striated pavements, geomorphic features and tillites, only erratic bearing deposits, if interpreted as ice-rafted dropstones, provide direct evidence of large-scale terrestrial ice. It is shown here that such erratics can be equally well explained by non-glacial processes such as organic-rafting. Erratics in the Chalk of southern England and the middle Cretaceous of Australia are representative of the problems of "glacial" indicators in the Mesozoic.

1. This chapter was written in collaboration with David Rowley (University of Chicago; Markwick and Rowley, In Press); consequently, throughout this chapter the plural pronoun "we" is used instead of the singular "I.".

The interpretation of high frequency, large magnitude, eustatic sea-level oscillations requires the existence of large ice sheets throughout much of the Mesozoic and Cenozoic. Although a temporal and spatial relationship between erratic-bearing deposits and transgressions is noted, the lack of demonstrable evidence for glaciation suggests that the amplitude of interpreted sea-level changes may have been exaggerated. We suggest that during non-glacial times the magnitude of any glacio-eustatic signal would have been no greater than 10-15 m, and generally much less. Such oscillations would have been swamped by intra-basinal effects. We therefore suggest that global correlation of derived "3rd-order" or equivalent eustatic curves is untenable. We further suggest that beach front erosion of forested coastlines during sea-level rises (local or eustatic) may better explain the origin of erratic-bearing deposits than continental ice-sheets and icebergs.

II.1. INTRODUCTION

Since the early nineteenth century, the general perception of Mesozoic and early Cenozoic climates has been of global warmth and ice-free poles (Beaumont, 1836; Frakes, 1979; Lyell, 1830). But recent work has suggested that this perception may be in error and that an ice-free Earth may never have existed (Frakes and Francis, 1988). This suggestion, however, is not new. Mesozoic ice has been proposed since the discovery of erratics in the Cretaceous Chalk of southern England (Mantell, 1833). Indeed, the climatic paradox that such erratics created led to a considerable literature on the problem throughout the nineteenth century (Croll, 1875; Godwin-Austin, 1858, 1860; Lyell, 1841, 1872; Martin, 1897; Stebbing, 1897), most of which still has relevance today. It is such erratics that still form the crux of arguments for the existence of Mesozoic-early Cenozoic glaciation (Frakes and Francis, 1988).

Arguments concerning the presence or absence of Mesozoic-early Cenozoic ice-sheets also have major implications for eustacy and thus the potential use of sequence stratigraphy for global correlation. Fluctuations in global sea-level (eustacy) require changes in the volume of the ocean basins and/or the volume of sea-water. As has been recognized by many workers, repeated, large-amplitude, high frequency eustatic oscillations, such as those implied by the 3rd order cycles of Vail et al. (1977) and the "short-term curve" of Haq et al. (1987) can presently only be explained by glacio-eustatic processes (Haq, 1991; Pitman, 1978; Pitman and Golovchenko, 1983; Rowley and Markwick, 1992). This is shown clearly in Table II.1 which tabulates the mechanisms and estimated rates by which both ocean basin volume and ocean water volume can change.

Rowley and Markwick (1992) derived a relationship for the volume change required for a specified change in sea-level, assuming that the -200 m to +300 m hypsometry (Cogley, 1985) has not changed significantly over time:

$$V_{sw} = 0.5\ A\ dSL + A_0\ dSL \qquad (1)$$

where V_{sw} is the volume of sequestered water, dSL is the change in sea-level from the present, A_0 is the initial surface area of the ocean, and A is the additional area due to dSL. The results are shown in Figure II.1, which also gives the ice coverage required if the water is sequestered as ice. Rowley and Markwick (1992) applied this method to the eustatic 3rd order curves of Haq et al. (1987) using the first order, or "M," curve as the no-ice tectonic eustatic signal (Figure II.2). Because high stands on the "M" curve represent the flooding of large areas, A is large and the volume of sequestered water necessary for a specified sea-level change is correspondingly greater than were the same sea-level change

TABLE II.1. MECHANISMS CONTROLLING SHORT AND LONG TERM EUSTATIC SEA LEVEL*

	Magnitude (m)	Duration (Myr)	Rate m/Myr
CLIMATE CONTROLS: OCEAN WATER VOLUME			
Glacial	120	20×10^{-3}	6000
Present Day sea level equivalents in modern ice masses (not corrected for isostatic effects)			
Antarctica†	74.1	-	-
East Antarctica†	65.5	-	-
Greenland§	6.6	-	-
Mountain Glaciers§	0.5	-	-
Dessication	10	3×10^{-3}	3300
Ocean temperature	10	1×10^{-3}	10000
TECTONIC CONTROLS: OCEAN BASIN VOLUME			
Long term ridge volume (maximum)	300	75	4
Long term ridge volume (mean)	200	75	2.7
Long term ridge volume (minimum)	100	75	1.3
Maximum interval ridge volume	45	5	9
Collisional decrease in continental area	70	50	1.4
Extensional increase in continental area	15	70	0.2
Sediment infilling	35	50	0.7
Hot spots	35	70	5

* based primarily on Table 2 in Rowley and Markwick (1992)
† based on volume of ice given in Drewry (1983)
§ based on volume of ice given in Flint (1971)

to occur during periods of low stand on the "M" curve. Consequently, for major eustatic changes such as that of the late Turonian (Haq et al., 1987) large ice-sheets are required (Figure II.3). Thus, while evidence of glaciations exists for the late Paleozoic, as well as other definable geologic periods (Figure II.4), the absence of unambiguous evidence for Mesozoic-early Cenozoic ice-sheets is problematic (Haq, 1991). This has led many workers to question the validity of high frequency eustatic curves, especially of the magnitude of the 3rd order Haq curves (Carter et al., 1991; Christie-Blick et al., 1989; Rowley and Markwick, 1992). However, other workers have suggested that Mesozoic ice-sheets existed and that the evidence has been either misinterpreted or lost (Brandt, 1986; Frakes and Francis, 1988, 1990; Kemper, 1983, 1987).

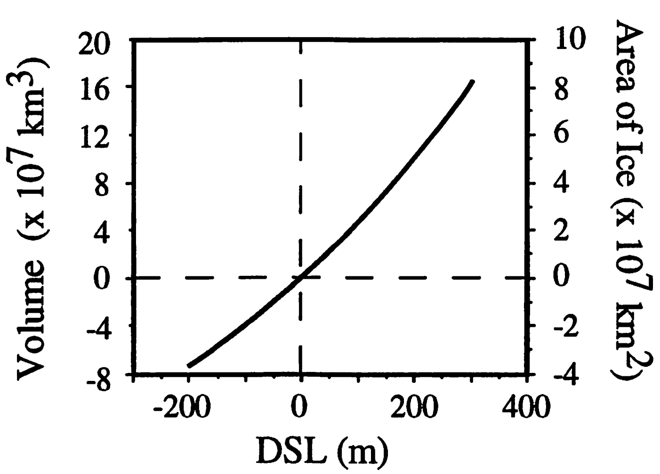


FIGURE II.1. The relationship between changes in global sea-level (dSL), volume of water sequestered and ice area required.

In this study we examine the evidence for Mesozoic-early Cenozoic glaciation using published reports dating back to the early nineteenth century. We also consider the literature on rafting mechanisms and provide calculations of carrying capacities of wood to assess its potential as a rafting agent in the geological record. Information on past glaciations is principally taken from the compilation of Hambrey and Harland (1981), which has been entered into a computerized relational database. The integration of this information with a study of glacial sedimentation and our compilations and analysis of the Mesozoic and Cenozoic paleoclimatic record enables us to assess the confidence with which we can consider glacio-eustatic sea-level changes, and also to understand the overall history of global climate change. We have also attempted to clarify misconceptions that have become embedded in the geological literature such as the definition of a "glacial interval" and our general perception of global paleoclimate. It is these questions that we address first.

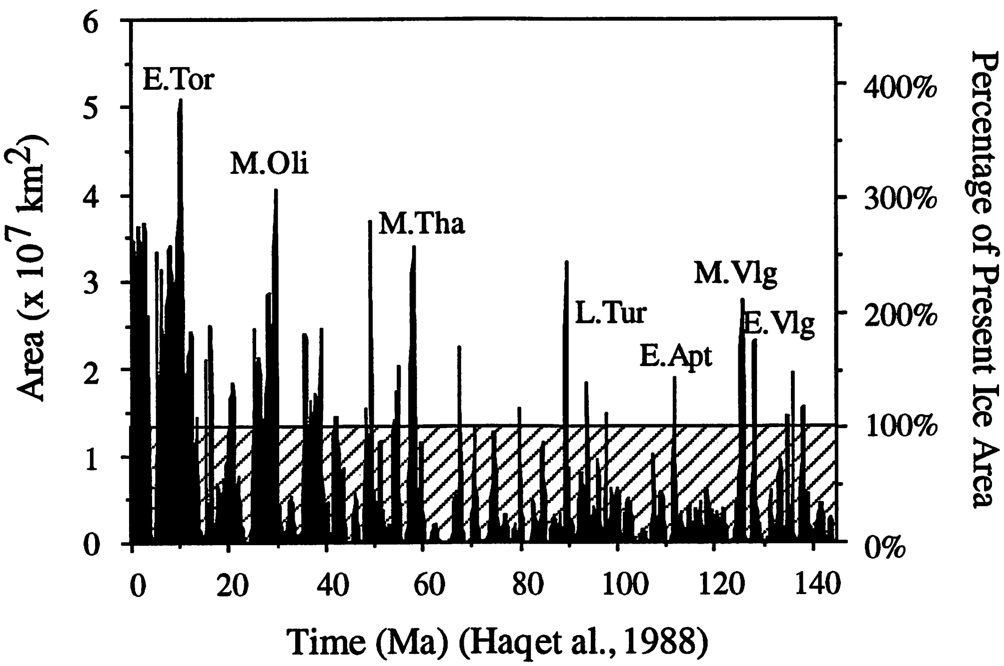


FIGURE II.2 The area of ice implied by the 3rd-order eustatic sea-level curve of Haq et al. (1987). After Rowley & Markwick (1992).

II.2. WHAT IS A GLACIAL INTERVAL?

In this study we define a "glacial interval" as "an interval with glaciation" using Gary et al.'s (p.299, 1972) definition of a "glaciation" as the "covering of large land areas

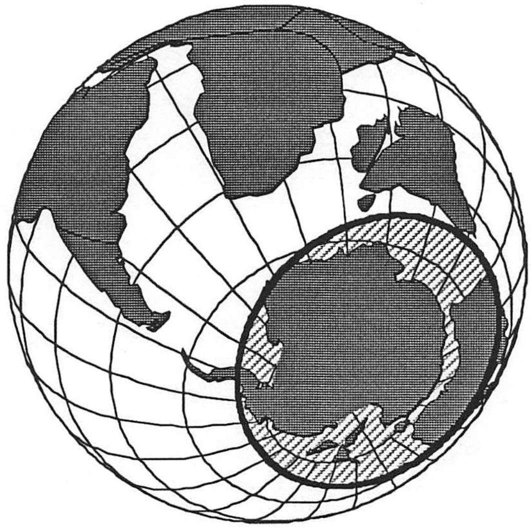


FIGURE II.3. The area of ice required to account for the Upper Turonian sea-level fall of Haq et al. (1987).

All of Antarctica and Australia would need to be covered. Such ice sheets would not extend into the deep oceans and this residual would have to be placed elsewhere, most probably in the northern hemisphere.

by glaciers or ice sheets." Implicit in this definition is that large volumes of water must be sequestered from the hydrological cycle, which will affect eustacy. As shown in Figure II.1, even small changes in global sea-level require the development of substantial ice sheets, such as the 6.6 m change associated with the deglaciation of Greenland (1.7 x 10^6 km^2) or the 74.1 m change associated with Antarctic (12 x 10^6 km^2) deglaciation. An important consequence of this definition is that it does not preclude the existence of some ice during "non-glacial" periods, nor does it require global climatic "equability" or high latitude warmth. As noted by North and Crowley (p.482, 1985) "*...an ice free earth does not necessarily imply a particularly warm earth.*". This must be stressed. Generalizing global climate, such as "warm Cretaceous," is extremely misleading: low latitudes, even during glacial times such as today, were probably always "warm," while high elevations in

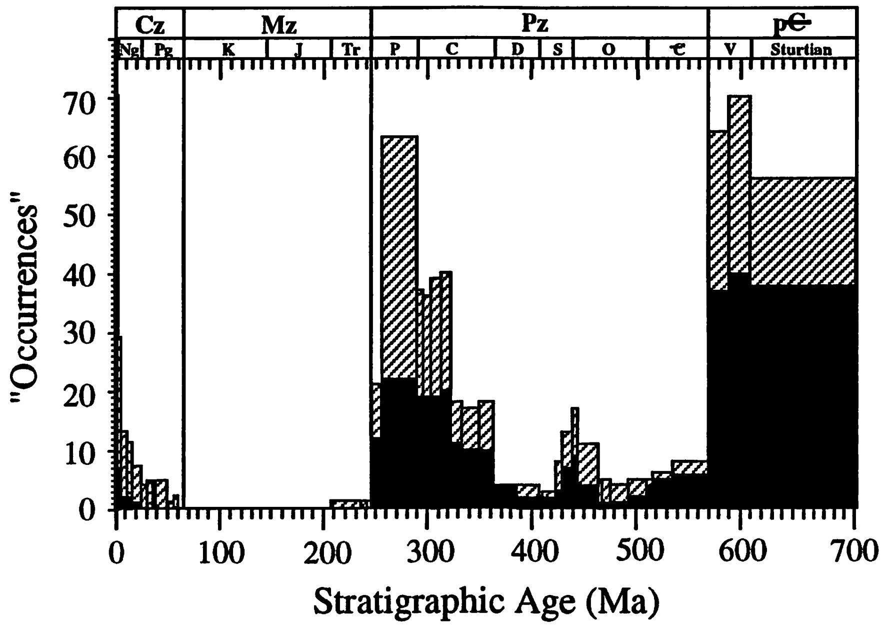


FIGURE II.4. Distribution of glaciation in the geologic record as implied by the frequency of glacial diamictite occurrences.

Diagonally shaded bars represent glacial diamictites, black shaded bars represent glacial diamictites with striated and faceted clasts. Data are taken ostensibly from Hambrey and Harland (1981).

high latitudes were probably always "cool" or "cold." Similarly, the practice of applying the paleoclimatic intepretations of a single locality to the entire world is at best naive.

II.2.1. Essentials for Creating a Glacier

The presence of glaciers is dictated by the intersection of the Equilibrium Line Altitude (ELA, the line on a glacier where net accumulation of ice and net ablation - ice loss - are equal) and the Earth's topographic surface (Figure II.5). The exact shape of the ELA is dependent on temperature, available moisture, and solar radiation (Charlesworth, 1957; Seltzer, 1994). Given sufficient elevation, glaciers can occur even in equatorial regions, as they do today in east Africa and Papua New Guinea. However, as the ELA rises, the

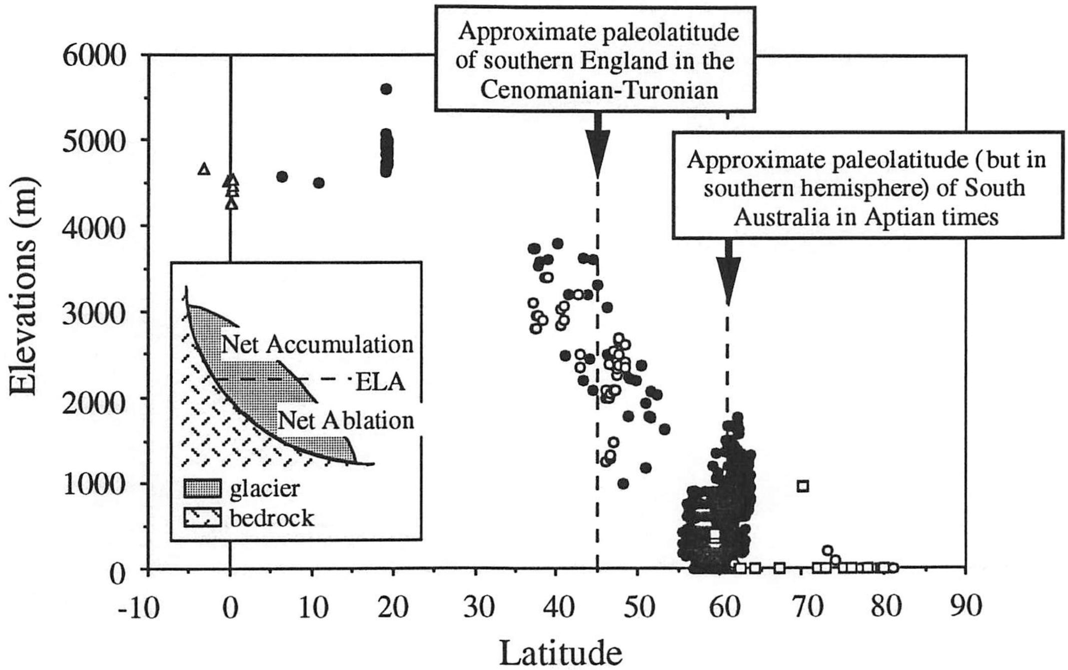
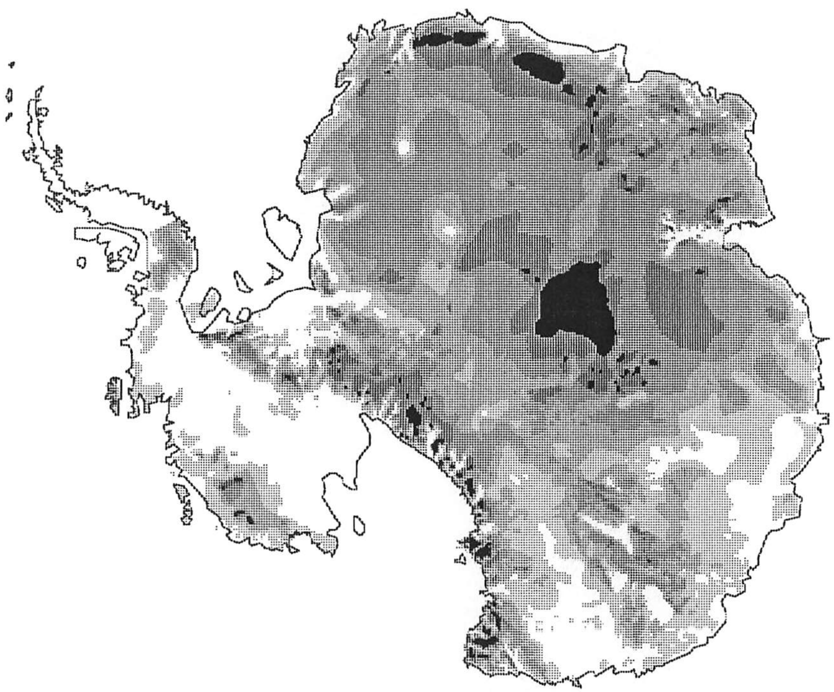


FIGURE II.5. The elevation of glacier termini versus latitude.

Black filled circles = American Cordillera glaciers; open circles = European glaciers; open triangles = East African glaciers; open squares = Canadian Arctic glaciers. Data are from the following sources: Field 1975; Osmaston 1989a,1989b. Insert shows the relation of the Equilibrium Line Altitude (ELA) to net ice accumulation and ablation (ice loss).

potential area for accumulation decreases and consequent glacial volumes are small with glacial terminations well above sea-level (Figure II.5). Consequently, mountain glaciers, as distinct from ice sheets, comprise only ≈0.6% of the total ice sequested water today, which is equivalent to about a 0.3 m sea-level change once water loading effects have been considered. During the last glacial maximum when the ELA was ≈900-1000 m below its present value irrespective of latitude (Pelto, 1992; Robin, 1988; Seltzer, 1994), mountain glaciers accounted for ≈1.5% of ice sequestered water (Flint, 1971). Thus the presence of mountain glaciers in the geological past can be assumed to have had a negligible affect on eustacy.

For the cryosphere (the portion of the Earth represented by ice) to have a significant effect on eustacy (i.e., >5-10 m sea-level change), large continental ice-sheets must develop. This requires that accumulation must be greater than ablation for large areas of the world. For this to occur there must be a significant depression of the ELA and/or the topography must rise. Robin (1988) calculated paleo-ELA's, despite the inherent problems involved (Andrews, 1975; Flint, 1971), and estimated that during the middle Cretaceous (100 Ma) the ELA was at ≈2000 m at the poles and ≈2200 m at 75° paleolatitude. Based on a quantitative analysis of the fossil floral data of the North Slope of Alaska, Spicer and Parrish (1990) suggested permanent ice existed at elevations above 1700 m at 75°N in the Cenomanian. Using these values we can estimate the size of the ice-sheet that would form by assuming the following: Drewry's isostatically reconstructed ice-free topography of Antarctica (Drewry, 1983; see Figure II.6) represents a plausible pre-glacial geography, and the ELA divides a glacier into two equal parts. With an ELA of 2000 m, $\approx 1.4 \times 10^6$ km^2 would be covered with ice, equivalent to a sea-level change from present of only ≈3.7 m. During the Cretaceous this effect would have been much less due to the higher sea-level. Even were the ELA to have been as low as 1500 m, the sea-level effect would have been no more than ≈15.6 m. The absence of glacial evidence on the continental margin off Dronning Maud Land, which represents part of this pre-glacial elevation, suggests that even such "small" ice-sheets did not exist during this period. The effects of higher ELA's during "warmer" geological periods would have been compounded or perhaps even induced during times when global topography itself (especially in high latitudes) was lower than present. This is a scenario implied for much of the Mesozoic (Figure II.7). However, with ≈97% of Antarctica presently hidden beneath ice any geological reconstructions of that continent's glacial and pre-glacial evolution must remain speculative.



Elevation (m)	Area ($x10^6$ km^2)	dSL (m)
> 2000	0.3	1.9
2000–1500	1.4	7.8
1500–1000	4.9	25.0
1000–500	7.5	38.6
500–0	9.8	50.7

FIGURE II.6. The ice-free, isostatically corrected, topography of Antarctica (after Drewry, 1983).

The area refers to the amount of land above each elevation, dSL is the sea-level change from present due to an ice-sheet covering that area.

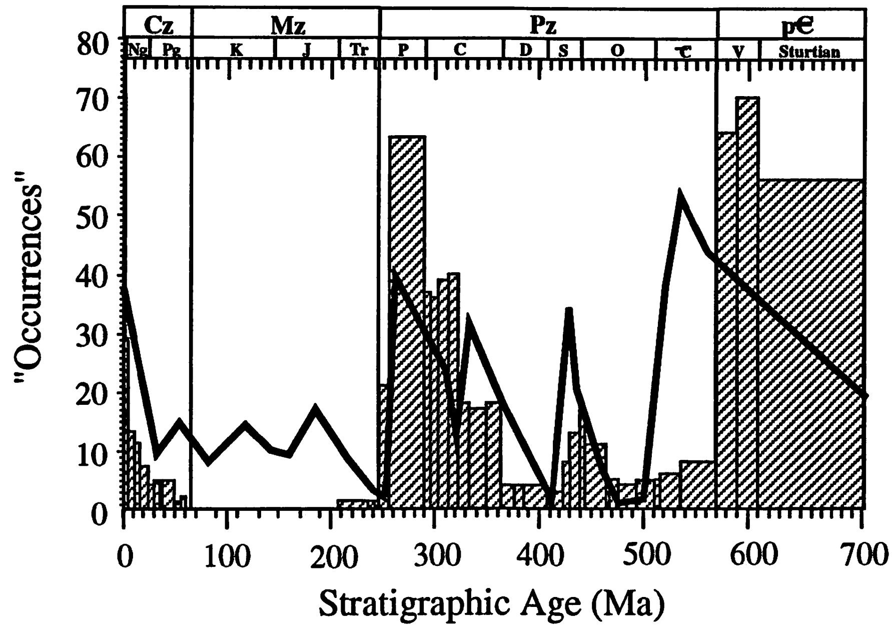


FIGURE II.7. The distribution of glacial diamictite occurrences through time on which is superimposed the area of deformation as derived by Richter et al. (1992).

Shaded bars represent glacial diamictites, thick black line represents the relative area of deformation. Note the correspondence between periods of glaciation and large areas of deformation. Deformation is used here as a proxy for mountain building.

II.3. PUTATIVE GEOLOGICAL EVIDENCE FOR MESOZOIC AND EARLY CENOZOIC GLACIATIONS

Most of the evidence for past glaciations is sedimentological and includes the presence of tillites (diamictites formed by glacial processes and usually evidenced by associated striated clasts and other glacial indicators), striated pavements and paleo-geomorphological features such as "U"-shaped valleys. This sedimentological evidence has been reviewed by numerous other workers (Boulton and Deynoux, 1981; Brodzikowski and Loon, 1991; Dowdeswell and Scourse, 1990; Eyles, 1993; Hambrey and Harland, 1981). The more salient features are summarized in Table II.2 based on information in Hambrey and Harland (1981). Although such features as striated pavements and striated and faceted clasts have been recognized in other definable "glacial" geological periods (Figure II.1), this is not the case for the Mesozoic and early Cenozoic where the sedimentological evidence is ambiguous at best and restricted to two phenomena: cyclicity, interpreted as due to glacio-eustatic oscillations (this also includes eustatic curves such as the 3rd-order curves of Haq et al. 1987), and pebbly mudstones, interpreteted as ice-rafted dropstones. Oxygen isotopic data provides the only other direct method for retrodicting past ice volumes (Matthews, 1984). Other "corroborating" evidence such as glendonites and stable isotope paleo-thermometry imply only the existence of cold conditions and not explicitly the presence of large ice-sheets. Paleontology, the source of most paleoclimatic data, is little help in this case since there are no fossils that are direct indicators of ice-sheets. Thus, the use of faunal changes, low diversity, the presence of "cold-climate" assemblages, or the absence of "warm-climate" organisms is purely circumstantial.

Therefore great emphasis is placed on the interpretation of sedimentological and geochemical evidence, and although we do discuss glendonites, the emphasis of this study is to test the validity of the sedimentary evidence for Mesozoic-early Cenzoic glaciation, specifically erratic bearing deposits, and to similarly assess the implications of the isotopic record.

II.3.1.Rhythmites, Cyclicity, and Eustatic Sea-level Change

Rhythmites (repeated lithological couplets or sequences) are common in the geological record. Although such sequences may be formed by changes in productivity, sedimentation rate, ocean chemistry or dissolution, many have been recognized to represent

TABLE II.2. INDICATORS OF GLACIATION

INDICATOR	DESCRIPTION	COMMENTS
Abraded bed rock surfaces (striated pavements) TERRESTRIAL	presence of striations, grooves, polished pavements, chattermarks, and rouche moutonneé like forms	although striated surfaces could be formed by tectonics, impacts, or mass-flow, generally such features are diagnostic.
Depositional fossil landforms TERRESTRIAL	e.g. drumlins, moraines, eskers, ice-push features, U-shaped valleys	rarely preserved
Chattermark trails on garnets TERRESTRIAL & SUBAQUEOUS	present in glacial deposits not in non-glacial deposits	requires garnets be present and preserved
striated, faceted (soled) and chattermarked clasts TERRESTRIAL & SUBAQUEOUS	striations on pebble surfaces generally in diamictites but also as isolated dropstones in marine settings. Faceting due to subglacier abrasion	can be caused by tectonic causes, for instance in shear zones , impacts, or by abrasion in mass-flow deposits. However generally one of the more diagnostic features associated with glacial sediments.
Stone-rich beds (diamictites) TERRESTRIAL & SUBAQUEOUS	unsorted, large range of grain sizes, may have preferred orientations, irregular thicknesses possible due to filling of hollows etc.	most of these features can be true for mass-flow deposits. Both can also have slump structures.
Dropstones SUBAQUEOUS	best observed in fine laminites where they show disruption of laminae at bottom and draping over top of stones. May be associated with graded laminites (varves). Stones may be striated, soled. Stones must be larger than matrix.	can be dropped from rafts other than ice, including kelp, trees, marine vertebrates and large birds. Certain mass-flow deposits may resemble dropstone-deposits but the disruption of laminites should be diagnostic of rafting.
'Fragile stones' TERRESTRIAL & SUBAQUEOUS	shale fragments mixed with more resistant material. Indicative of rapid deposition and lack of water transport.	mass flow deposits might have same

TABLE II.2., continued

INDICATOR	DESCRIPTION	COMMENTS
Variable lithologies TERRESTRIAL & SUBAQUEOUS	clasts are exotic (may be far traveled), and have variable lithologies dependent on size of glacier/ ice sheet and bedrock overridden. Minerals often undecomposed suggesting little transport by water and cool conditions resulting in reduced chemical weathering.	dropstones from tree roots or kelp may also show a variety of lithologies depending on source of material; however because of the large amount of chemical weathering associated with organics and soils such material should show more decomposition of unstable minerals (e.g. feldspars)
Stones with wide range of shapes TERRESTRIAL & SUBAQUEOUS	well-rounded to angular, especially sub-angular and sub-rounded	some mass flow deposits may show similar variations as might clasts due to tree rafting etc., depending on source of clasts; however this is quite a good indicator.
Quartz grain textures TERRESTRIAL & SUBAQUEOUS	angular grains, sharp edges and conchoidal fracture	such textures can be subsequently altered
Large areal extent and thickness TERRESTRIAL & SUBAQUEOUS	unsorted deposits over large distances (km's)	mass-flow deposits may be more localized
Finely graded stratification SUBAQUEOUS	laminated clay or siltstones	also found in distal turbidites and some storm deposits
Rock flour TERRESTRIAL & SUBAQUEOUS	clay sized particles ground by the action of the moving glacier.	such particles are indistinguishable under the microscope
Proxy Data (evidence of cooling and cold conditions)	includes fossil data, cold water assemblages, low diversity, etc. glendonites	cold does not explicitly mean glaciation

cyclic shallowing-deepening oscillations. Where such oscillations are demonstrably eustatic, high amplitude (>5-10 m) and rapid (>100 m/myr), a glacio-eustatic origin is the only explanation (Rowley and Markwick, 1992). Indeed, Bjørlykke (1985) suggested using sea-level changes in tectonically stable areas to constrain the timing and magnitude of ancient continental glaciations where such changes are globally correlative (eustatic). In most cases this is not clear, but some workers have chosen to use such evidence. Brandt (1986), for example, interpreted glacio-eustatic cycles in the Early Jurassic of Germany based primarily on the contemporary occurrence of glendonites and pebbly mudstones. Kemper (1983) interpreted an Early Cretaceous regression in northern Russia as due to "isostatic uplift of the Arctic as a consequence of melting of a thick Arctic ice sheet", despite the lack of any corroborating evidence of ice. In addition, a regression in northwestern Europe during Plenus Marl time (Cenomanian-Turonian) was interpreted as glacio-eustatic by Jeans et al. (1991), based on isotopic data and the presence of erratics, erratics that in fact also occur before and after the "glaciation." Most important is the Haq 3rd order curve which, as shown by Rowley and Markwick (1992) and acknowledged by Haq (1991), implies the existence of large ice-sheets since at least the Kimmeridgian that would have to have been larger than the present Antarctic ice sheet on at least 25 occasions (Figure II.2).

II.3.2. Pebbly Mudstones and Erratics.

Erratic bearing deposits, including pebbly mudstones (a rock comprised of dispersed pebbles in a mudstone matrix; Crowell, 1957), present an immediate depositional paradox since they imply two very different hydrological regimes (the term "erratic" is used throughout this text according to the definition of Gary et al., 1972, p.238): a "*relatively large rock fragment lithologically different from the bedrock on which it lies, either free or* *as part of the sediment, and that has been transported, sometimes over a considerable distance, from its place of origin.*" Unfortunately, for most geologists the term also has glacial connotations, which we do not imply; however other terms such as "megaclast" or "lonestone" are less appropriate since they do not convey the exotic nature of the clasts. Although mass-flow may account for some of these deposits (Crowell, 1957), and may be responsible for reworking others, a mass-flow origin is not always demonstrable, and so rafting mechanisms, particularly ice, have often been invoked. In the Mesozoic such deposits have drawn particular attention because of the lack of more direct evidence for glaciation. There are many erratic-bearing marine deposits in the Mesozoic and early Tertiary, a selection of which is presented in Table II.3. Two erratic bearing deposits have received particular attention: the Chalk of northwestern Europe, and the middle Cretaceous rocks of Australia. It is the erratics of these two units that presently provide the crux of arguments for Mesozoic glaciation (Frakes and Francis, 1988; Jeans et al., 1991). If such erratics are not demonstrably due to ice-rafting, then such arguments become moot. We thus first discuss the erratic bearing deposits of these two areas, then review erratic emplacement mechanisms so as to better constrain plausible interpretations.

TABLE II.3. SUMMARY OF PRE-PLEISTOCENE ERRATIC-BEARING DEPOSITS

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
Maruyama Formation Sakhalin (Pliocene)	quartzites, chert, andesites, schists, silicified mudstones, sandstones, basaltic porphyries, siliceous mudstones, clay balls	usually 1 - 5 up to 1500 - 2000	usually rounded, occasionally angular		45, 47
Kakert Formation Kamchatka (Miocene)	<u>volcanics</u>, vein rocks, sedimentary rocks	mostly 1 - 5	rounded to well-rounded	comprise ≈1-3% of rock; derived from underlying deposits; no striations or faceting known. Interbedded with tuffaceous sediments.	45, 46
Mackenzie Bay Formation Mackenzie Bay area (Miocene)	<u>chert</u>	pebbles	-	scattered	2
Utkholok Formation Kamchatka (Oligocene)	<u>volcanics</u> (45-50%) and <u>sedimentary</u> (40-45%) rocks, minor metamorphics, vein rocks, intrusive igneous, wood and coal	generally 0.1 - 1 some 20-30 rare up to 1500	rounded - well-rounded (very largest clasts are sub-rounded)	locally derived, all types present in underlying deposits, wood present . No striations or facets known. These beds are interbedded with coals and tuffaceous sediments	48
Takaradai Formation Sakhalin Island (late Eocene)	<u>siliceous rocks</u> and <u>quartz</u>, minor crystalline schists, quartzites, andesites, and shales	usually 1 - 5 up to 10	well-rounded, rarely angular	interbedded with intermittent coals and tuffaceous deposits	49

TABLE II.3., continued.

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
Gilsonryggen Formation Svalbard (?Eocene)	quartz, quartzite, chert, dolerite, gray granite, graphic granite, rhyolite porphyry, vein quartz, sandstone	quartz and chert are 1 - 5, dolerite blocks up to 15-50, porphyry pebbles are up 2 - 30	well-rounded; dolerite blocks are angular-sub-rounded	calcite rosettes are found (possibly glendonites?)	41, 42
Sarkofagen Formation Svalbard (Eocene)	95% comprise <u>chert</u> and <u>quartz/quartzite</u>; rarer gneiss, granite, greenschists, sandstones, volcanics	granules and pebbles, some rare cobbles and boulders	generally very well-rounded	restricted to marine parts of sequence. clasts occur in 3 settings: scattered, concentrated on hummocky erosion surfaces, as horizontal concentrations, the lower boundary often been sharp. fragments of fossil wood are known form this formation. calcite rosettes are found (possibly glendonites?)	40, 42
Pukemuri Formation New Zealand (early Eocene)	igneous rocks, greywacke, green sandstone	2 - 30	mostly rounded, but some are sub-angular	in sandy - silty matrix, large scattered boulders in basal part of formation, clasts generally derived from immediately underlying deposits	38
Sweden (Danian)	black and white <u>cherts</u>	2.3 - 7.1cm long	rounded, smoothed	associated with crocodilian remains (gastroliths?)	37

TABLE II.3., continued.

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
Denmark (Danian)	quartz, schist, gneiss, granite, rhyolite, gabbro, sandstone, shale	up to 7 x 5.5 x 3.5	rounded - sub-angular	found in coral limestones and bryozoan limestone, calcarenites	37, 43
Denmark (Maastrichtian)	<u>quartz</u>, quartzite, sandstone	up to 6.4 x 4.5 x 2.0	rounded - sub-angular	in chalk	43
Bearpaw Shale Montana (Campanian)	<u>black chert</u>	up to 1	rounded, smooth	scattered throughout shale	pers. obs.
Judith River Formation Montana (Campanian)	<u>chert</u>	up to 2	rounded, smooth	occur on revinement surfaces as lags during transgressions, associated with plesiosaur bones, shark teeth, mososaur bones & other vertebrate remains	35
Hue Shale NE Alaska (Campanian)	<u>chert</u> and <u>quartzite</u>	pebbles up to 4	-	floating in shale, common. organic rich (commonly >4%C_{org})	3
West Siberian lowland (lower Paleocene)	-	-	-	pebbly mudstone	45

TABLE II.3., continued.

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
'Piripauan beds' New Zealand (Campanian)	plutonic, volcanic, quartzite, sedimentary rocks	2 - 50	sub-angular	No striations	38
Upper Chalk S.England and NE France (Turonian-Maastrichtian)	quartz, quartz schist, quartz phyllite, quartzite, black chert, coal (Lydden Tunnel)	up to 300g, most around 2 - 8g, coal = 120 x 120 and 10 - 25 thick	mostly well-rounded, some angular	wood impressions and fish found in vicinity; nature of coal remains uncertain	15, 16, 17, 18
Middle Chalk S. England (Turonian)	<u>granite</u>, <u>quartzite</u>, greenstone, siliceous sand, sandstones, dark clay slate	up to 30.5 (Purley, UK)	"water-worn," well rounded	fossil wood found in vicinity of some, many boulders encrusted by serpulids, bryozoans and spondylids	17, 18, 19, 20, 21, 22, 23
Soester Grünsand Germany (Turonian)	gneiss, granite, monzonite, quartzite, greensand, chert	up to 45	rounded	isolated	24
Lower Chalk S. England (Cenomanian)	<u>quartzite</u>, cherts, vein-quartz, clay slate, "tourmalinised killas," sphaerulite, greenstone	"pebble," "large blocks," generally as small pebbles	well-rounded	stones mixed with rotten wood (?) at Isleham; at Rochester 107 pebbles closely associated together	17, 19, 20, 21, 25, 26, 27
Osbourne Formation Western Australia (Albian-Cenomanian)	quartz	pebbles		in claystones	56

TABLE II.3., continued.

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
Upper Greensand S.England (upper Albian)	coarse feldspathic grit, purple shale, gray sandstone, fine conglomerate, black limestone, red sandstone, mica-schist, phosphate nodules, granite, greenstone, mica-hornblende-schist, talcose schist, vein quartz, quartzites, spherulitic rhyolite, labradorite	sizes range from 5 - 55 up to 35.5 x 30.5 x 15	angular - rounded, generally angular (c.10% rounded)	some coated in phosphate, many encrusted by bivalves, faint striations on one phosphate nodule (striations then covered by Cretaceous encrusters), faint striations on one mica schist (?); some specimens seem highly weathered others are very fresh (no decomposition)	17, 28, 29, 30
Gault Clay S. England (Albian)	quartzite, milky quartz, black slate, hornstone, sandstone	boulders (25.5 x 15 x 15 and 18 x 18 x 15), mostly pebbles	boulders are angular, but generally pebbles are rounded (some "water-worn"). Flat polished pebble found near Brighton	boulders encrusted by *Plicatula* and *Spondylus*. At Folkstone pebbles associated with remains of a plesiosaur (gastroliths?)	25, 31, 32, 33
Manahau Formation New Zealand (Albian)	rhyolite, ignimbrite, tuff, plutonic rocks, chert	2 - 5	well-rounded, polished, scattered	clasts are exactly the same as those of the immediately underlying conglomerates	38

TABLE II.3., continued

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
Carolinefjellet Formation (Innkjegla Member) Svalbard (upper Aptian - lower Albian)	chert and quartzite	pebbles	-	driftwood also found in this member	39
Bulldog Shale South Australia & New South Wales (Aptian)	Devonian quartzites, volcanic rocks	boulders, up to 300	well rounded		57, 58
Marree Formation South Australia (Aptian)	quartzite and rare shale	up to 50			62
Wallumbilla Formation Queensland (Aptian)	quartz	pebbles	well-rounded	associated with wood and vegetal matter	59, 60
Rumbalara Shale Northern Territory, Australia (Aptian)	quartz	boulders	-	silicified wood in this unit	

TABLE II.3., continued

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
Rolling Downs Formation Queensland and South Australia (Aptian)	<u>quartzite</u>, feldspar porphyry, granite, schists, gneisses	pebbles, boulders, mostly less than 61 but some up to 167.5 in length	sub-rounded - well-rounded ("water-worn")	associated with numerous fossil trees in South Australia	60, 61, 51
Monash Formation (Merreti Member) South Australia and New South Wales (Aptian)	lithic	pebbles and cobbles		in clays, slump structures	54
Monash Formation (Pyap Member) South Australia and New South Wales (Aptian)	-	cobbles and pebbles		in claystone interbeds, glauconitic, with carbonaceous material and pyritized wood fragments	54
Unit 6 (Mullaman Beds) Northern Territory, Australia (Aptian)	-	pebbles		associated with teredo-like bored wood	52
Cadni-Owie Formation South Australia (Neocomian-Aptian)	<u>quartzite</u>, rhyolite, others	pebbles, cobbles and boulders up to 120	sub-rounded - rounded		57, 62, 63

TABLE II.3., continued

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
Leederville Formation Western Australia (Neocomian-Aptian)	quartz	pebbles		in claystones	56
unit not given, probably the *Mullaman Beds* Northern Territory and Queensland (Neocomian-Aptian)	mostly sedimentary	up to 26, generally granules and small pebbles	rounded - sub-angular	no striae or chattermarks.	53
Blythesdale Sandstone South Australia (Neocomian-Aptian)	<u>quartzite</u>, quartz feldspar porphyries, quartz, slate, schists	up to 182	rounded	some faceted, no striations, wood and leaf impressions, associated fossil wood	50, 51
Loongana Sandstone South Australia (Neocomian-Aptian)	<u>granite</u>, schist,	>10 (the diameter of the core)		in feldspathic sandstone, some of the granite and schist pebbles may have been in a conglomerate	4
Unit 2 (Mullaman Beds) Northern Territory, Australia (Neocomian)	-	pebbles and cobbles	-	associated with sandstones and conglomerates which have poorly rounded pebbles and boulders. Clasts also concentrated in lenses. Very fossiliferous unit	36, 52

TABLE II.3., continued.

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
Pebble Shale North Slope, Alaska (Hauterivian-Barremian)	<u>chert</u>, <u>quartzite</u>	pebbles or granules with rare cobbles		floating sand grains and pebbles, pebbles are not concentrated (although there is some increase in number of clasts) in basal part of shale unit as is typical of most transgressive shale units. Unit is organic rich (possible oil source rock for Prudoe Bay oil-field)	34, 3, 36
Kemik Sandstone (Marsh Creek Member) NE Alaska (Hauterivian-Barremian) below Pebble Shale Unit	<u>chert</u>, <u>quartz</u>	pebbles	-	common in interbedded black mudstone & dark gray argillaceous siltstone	55
northern Siberia (Middle-Upper Jurassic)	-	pebbles	rounded and slightly flat	in non-bedded silty claystones	44, 45
northern Siberia (Bajocian)	-	cobbles and boulders up to 200 long	-	-	44

TABLE II.3., continued

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
Lower Beaufort Shales South Africa (Middle Triassic)	-	up to size of an "orange"	"water-worn" pebbles	often associated with masses of fossil wood "*...great masses of silicified wood and that a large number of water-worn stones were lying amongst the wood.*"	1
Snowdon Formation New Zealand (upper Scythian or early Anisian)	chert, granite, argillite, basalt, andesite, granophyre, gabbro	3 - 35 diameter	angular - rounded	spaced between 20cm and 1 m apart, in black siltstone matrix, underlain by a conglomeratic member	38

AUSTRALIAN ERRATICS: the relationship of the erratic bearing horizons is further confounded by the plethora of names used for the beds in which they occur. In this summary the names are those as given by the workers given in the reference column. To clarify , the following units are laterally equivalent: Wallumbilla Formation = Bulldog Shale = Roma Formation = Unit 6 (Mullaman Beds) = Rumbalara Shale = lower part of Maree Formation = Merreti Member of the Monash Formation = lower part of Rolling Downs Formation (Day, 1969; Exon and Senior, 1976). The Cadni-Owie Formation stratigraphically underlies these units and is equivalent to the Pyap Member of the Monash Formation (Thornton, 1974) and may be equivalent to the Leederville Formation of the Perth Basin , unit 2 of the Mullaman Beds and the Blythesdale Sandstone of South Australia (Day, 1969).

TABLE II.3., continued

Stratigraphic unit with age and general location	Lithology of erratics. Underlining infers dominant lithology(s)	Size (diameter, cm)	Shape	Association	References
FOR COMPARISON					
Carboniferous coals UK & Germany (Pennsylvanian)	<u>quartzite,</u> quartz, granite,	"small" (Germany); up to 11 x 8 x 3.8 (Wales); 2.5 - 20 (England); pebbles up to 25 (Derbyshire, UK)	rounded ("water-worn"), well-worn	one in situ in coal layer (Wales), in coal seam (Staffordshire, UK); enveloped in coal (Leicestershire, UK); in underclay (Derbyshire, UK)	5, 6, 7, 8, 9
Carboniferous limestones Ireland (Carboniferous)	<u>granite,</u> schist, quartzite, vein-quartz, slate	up to 20 diameter	angular	little evidence of decomposition of fragments	10
Carboniferous coals USA (Pennsylvanian)	<u>quartzite</u>, vein-quartz, sandstone conglomerate, granite, quartz porphyry, limestone, talcose slate, grit, phyllite, pegmatites	up to 48 x 38 x 35.5 (73.5 kg) (West Virginia); up to 60, most between 1 and 4 (upper Pottsville Fm, Alabama)	angular to well -rounded, mostly rounded and polished	some faceted clasts, some from within coal seams; in upper Pottsville Fm., clasts associated with wood remains, largest log is 700 long and 50 diameter	5, 12, 11, 13, 14
La Salle Limestone IL., USA (Pennsylvanian)	vein-quartz, greenstone, granite, schist, quartzite	up to 11	smooth, possible faceting	no striations	11

TABLE II.3., continued

REFERENCES

1 - Broom, 1911; **2** - Nentwich and Yole, 1982; **3** - Molenaar et al., 1988; **4** - Lowry, 1970; **5** - Price, 1932; **6** - Dix, 1944; **7** - Spencer, 1887; **8** - Bonney, 1873; **9** - Gresley, 1885; **10** - Ball, 1888; **11** - Savage and Griffin, 1928; **12** - Newberry, 1872; **13** - Hicks, 1879; **14** - Liu and Gastaldo, 1992; **15** - Cayeux, 1897; **16**- Robbie, 1950; **17** - Stebbing, 1897; **18** - Godwin-Austin, 1860; **19** - Double, 1931; **20** - Hawkes, 1951; **21** - Dixon, 1850; **22** - Dibley, 1910; **23** - Dibley, 1918; **24** - Schmidt and Schreyer, 1973; **25** - Jukes-Browne and Hill, 1903; **26** - Godwin-Austin, 1858; **27** - Strahan, 1906; **28** - Sollas and Jukes-Browne, 1873; **29** - Hawkes, 1943; **30** - Watts, 1881; **31** - Woods, 1895; 32 - Seeley, 1877; **33** - White, 1924; **34** - Molenaar, 1988; **35** - Rogers, 1994; **36** - Walpole et al., 1968; **37** - Troedsson, 1924; **38** - Waterhouse and Flood, 1981; **39** - Nagy, 1970; **40** - Dalland, 1977; **41** - Birkenmajer and Narebski, 1963; **42** - Birkenmajer et al., 1972; **43** - Noe-Nygaard, 1975; **44** - Chumakov, 1981 ; **45** - Epshteyn, 1978; **46** - Grechin, 1981a; **47** - Melankholina, 1981a; **48** - Grechin, 1981b; **49** - Melankholina, 1981b; **50** - Woodard, 1955; **51** - Woolnough and David, 1926; **52** - Skwarko, 1966; **53** - Frakes and Krassay, 1992; **54** - Thornton, 1974; **55** - Mull, 1987; **56** - Playford and others, 1976; **57** - Flint et al., 1980; **58** - Frakes and Francis, 1988; **59** - Moore, 1870; **60** - Jack and Etheridge, 1892; **61** - Jack, 1931; **62** - Carr et al. 1979; **63** - Wopfner et al. 1970

II.3.3. The Erratics of the Chalk and Upper Greensand of Southern England

Mesozoic erratics have been described since their discovery in the Chalk of southern England in the early nineteenth century (Mantell, 1833). The Chalk, inextricably linked with global warmth and aridity (Bailey, 1924; p.54, Chatwin, 1960; p.38, Gallois, 1965; Hancock and Scholle, 1975; Hancock, 1975; Rawson, 1992), seemed immediately to preclude ice rafting (Lyell, 1841). Similar erratics have been found in the underlying greensands (Hawkes, 1943), the Gault Clay (p.196 and 346, Jukes-Browne and Hill, 1903; Dines et al., 1933; White, 1924), the Soest Greensand of northern Germany (Kaever, 1974; Schmidt and Schreyer, 1973), the Chalk of the Paris Basin (Cayeux, 1897), the Danian and Maastrichtian limestones of Denmark (Noe-Nygaard, 1975) and Sweden (Troedsson, 1924), and the Mesozoic and Tertiary of Spitsbergen (Birkenmajer et al., 1972; Nagy, 1966, 1970; Woolnough and David, 1926). Most of the erratics found in the Chalk and Upper Greensand reside in museum and private collections and locality, stratigraphic and sedimentological details are generally scarce. Collection biases may have influenced both the spatial distribution of the erratics (Figure II.8; through mining operations, Double, 1931) and also size distributions from initial collection preferences for larger clasts. Throughout this discussion we use the traditional subdivision of the Chalk (Gallois, 1965).

The erratics of the Chalk comprise various lithologies, outlined in Table II.3, with granite (Godwin-Austin, 1858) and quartzites the most common types (Hawkes, 1951). With the possible exception of a large coal clast (the largest "erratic" reported from the

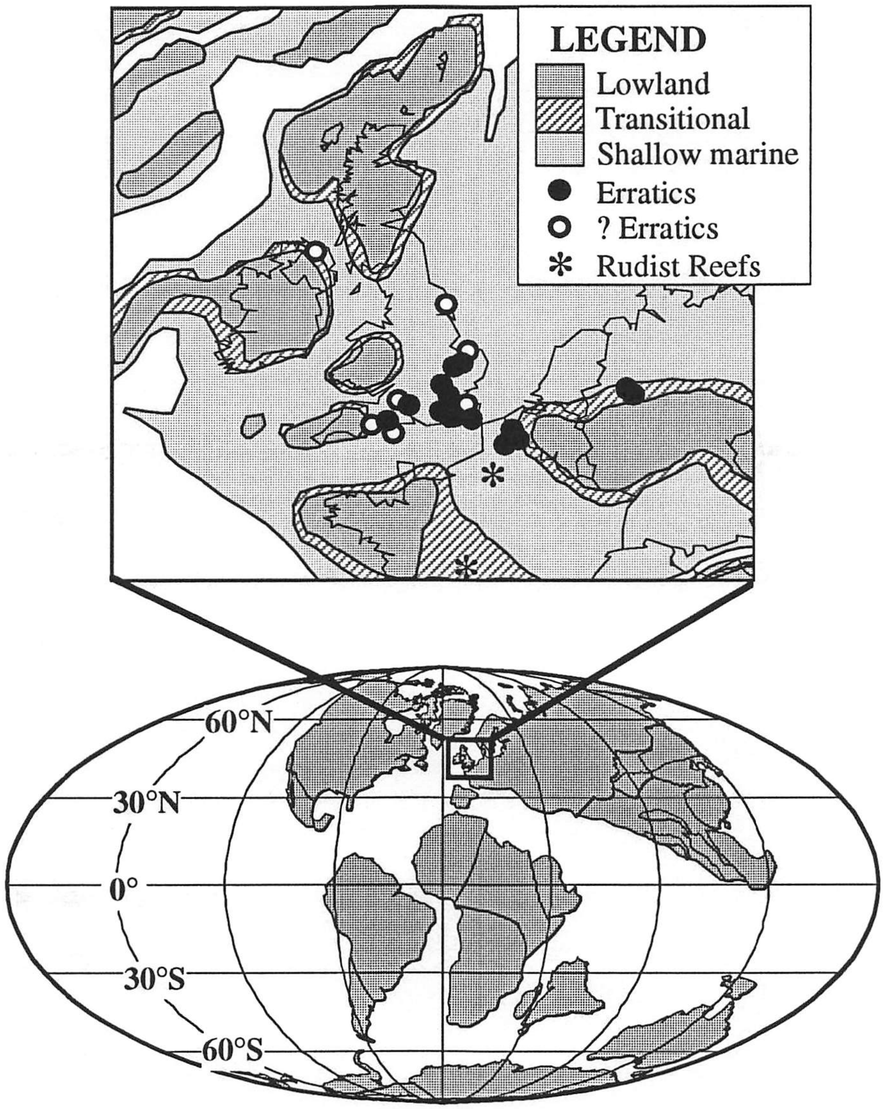


FIGURE II.8. Distribution of middle Cretaceous erratics in north-west Europe.

See Table II.3 for references. Paleogeography after Ziegler (1990). Plate reconstruction is for 92 Ma, Cenomanian (Rowley, unpublished data).

Chalk - 1.22 x 1.22 x 0.25 m and whose provenance is unclear, Godwin-Austin, 1860), all of the erratics in the Chalk are "water-worn," which is consistent with most of the purported dropstone deposits of the Mesozoic. The largest demonstrable erratic, known as the Purley boulder, was estimated to have been ≈ 91.4 cm in diameter (Woolnough and David, 1926), although the boulder's exact size is uncertain since it was broken up by the workers who found it (Godwin-Austin, 1858). Hawkes (1951) suggested a smaller size of 40.6 x 22.9 x 25.4 cm based on the fragments in the Geological Survey and Museum. Most of the pebbles in the Chalk are much smaller, and the size range of 0.5-5.0 cm quoted for most of those in the Wiltshire and Dibley collections is more typical (Hawkes, 1951). To date no faceted or unambiguously striated pebbles have been described.

Although generally found in isolation, associations of pebbles do occur, such as those at Rochester where 150 pebbles were found within a 1 or 2 m radius (Hawkes, 1951). At Purley, Godwin-Austin (1858) found that although the granite boulder had been removed, its original site did not seem "*...to have been much disturbed.*" Lyell (1872) added further that the boulder weighed over 18 kg. Associated with this material Godwin-Austin (1858) found a 9 kg boulder, "*..some coarse shingle, and a quantity of loose sea-sand.*" Betchworth quarry in Surrey, which was heavily used in the mid-nineteenth century, accounts for most of the described erratics in the Chalk (about 70 stones, Double, 1931). One granite cobble was found encrusted with several Cretaceous clams (lower valves of *Spondylus latus*) and serpulids (Double, 1931). Encrusted clasts have also been found in the Catt collection from the Lewes area, Sussex (fauna includes *Diblasus* or *Isis*, a serpulid, a bryozoan (*Diastopora ramosa*?) and a lower valve of *Spondylus lineatus*, Godwin-Austin, 1858), in the Upper Greensand of Cambridgeshire and the Gault Clay from Stanbridge, south Bedfordshire. Most of the erratics have been found from either the Lower or Middle Chalk.

The erratics of the Upper Greensand are also dominated by resistant siliceous lithologies. Hawkes (1943) noted that 94% of his samples were siliceous: mainly quartzites, greywackes and sandstones are present, with biotite-granite-gneiss, vein quartz, rhyolites and schists also quite common, with the remainder consisting of hornstone-tuffs, quartz-porphyries and more basic rocks. Compared with the erratics of the overlying Chalk, those of the Upper Greensand are generally more angular (Bonney, 1872; Sollas and Jukes-Browne, 1873), larger, and more common (Hawkes, 1943). They are usually found isolated (Bonney, 1872; Hawkes, 1943), although Sollas and Jukes-Brown (1873) report six "*large stones.......huddled together at Waterbeach.*" Except for possible wood associated with the pebbles at Isleham (Jukes-Browne and Hill, 1887; Stebbing, 1897), none of the pebbles is directly associated with organic matter. The Upper Greensand does, however, contain balls of vegetal matter (Bonney, 1872) as well as a diverse vertebrate fauna including crocodilians, turtles, fish and dinosaurs (Bonney, 1872; Chatwin, 1961).

II.3.3.1. Depositional Environment and Provenance

The Cambridge Greensand is a condensed deposit that formed on the northern margin of the London Platform and is the lateral equivalent of the Upper Greensand of the Weald. It seems to represent nearshore clastic shelf deposition (Chatwin, 1961), although the exact details are still unclear (Anderton et al., 1979). It is usually less than 1' (c.30.5 cm) thick, with "sandy and chalky marl with abundant phosphate nodules at the base" among which the erratics are generally found (Chatwin, 1961). The unit contains an abundant and diverse fauna including pterodactyls, dinosaurs, crocodilians and sharks, as well ammonites, bivalves, brachiopods and corals.

The Chalk is a fine grained (dominantly 4-10 μm), exceptionally pure limestone with greater than 96% $CaCO_3$ (Hancock, 1975). Although the Lower Chalk contains a large proportion of siliciclastic material, detrital material is generally extremely rare for most of the Chalk. However, it may become locally significant near presumed terrestrial sources: the upper Turonian Tuffeau Jaune de Touraune Formation in the southwestern part of the Anglo-Paris Basin has up to 20% angular quartz grains (Jarvis and Gale, 1984) and the Cenomanian Beer Head Limestone in southeast Devon has up to 50% quartz sand. In Northern Ireland, the White Limestone (Chalk equivalent) onlaps onto older rocks and also contains large amounts of detritus including pebbles of quartz, quartzite and quartz porphyry (Hume, 1897). The Chalk, like the Greensands below, contains an abundant and diverse fauna, although the strongly alkaline conditions seem to have largely precluded the preservation of plant material except within flints (Dixon, 1850). Hancock (1975) outlines the evidence supporting a soft "watery" nature for the Chalk sea floor. The bathymetric relief of the chalk sea bottom was not insignificant, with large carbonate banks, such as those of Normandy where the banks are up to 50 m high and 1.5 km long (Kennedy and Juignet, 1974) and of Portsdown, England (Gale, 1980). Slumps adjacent to these banks are common (Gale, 1980; Mortimore and Pomerol, 1991). Omission surfaces, scouring and hardground development are evidence of erosion and/or non-deposition (Kennedy and Garrison, 1975) and have been used to interprete sea-level changes during Chalk times. Hancock (1989) extensively discussed discrepancies between these changes and the eustatic curve of Haq et al. (1987).

The source of the erratics has been problematic. For those of southern England, much of the material was derived from older deposits (Hawkes, 1943), especially conglomerates such as those of the Triassic (Hume, 1897). Clasts of similar petrology were available in most of the underlying beds, such as the Upper Greensand, Lower

Greensand and especially the Wealden; provenance studies of these beds in turn show not only derivation from primary source areas such as Cornubia, Brittany or the London Platform (Allen, 1949, 1960, 1967, 1969, 1972, 1981; Kirkaldy, 1947; Sladen and Batten, 1984; Wells and Gossling, 1947), but also reworking from Liassic rocks (p.66, Edmonds et al., 1975). It is clear that cobbles and boulders were available on local uplifts such as the London Brabent Platform (Prestwich, 1856). Although source areas, as positive features, have probably been subsequently removed, onlapping directions and detrital compositions suggest possible sources. In the southwestern Paris Basin detrital quartz concentrations increase to the southwest (Jarvis and Gale, 1984); detrital concentrations in the Lower and Middle Chalk deposits of southern England generally increase westward towards Cornubia (Groves, 1931), concomitant with an increase in coarseness of the chalk. Smith (1961) suggested that the Cornubian Massif was submerged by the time of Middle Chalk deposition. Groves (1931) noted a similar trend for the Upper Greensand, particularly in East Devon, Wiltshire, and Lulworth Cove (the Upper Greensand of the Isle of Wight in Hampshire appears to be sourced from Brittany). Possible contemporary beach deposits have been reported in Devon (Ali, 1976; although Jarvis and Woodroof, 1984, dispute this), southern Belgium (p.300, Prestwich, 1888), and Northern Ireland (Hancock, 1961; Wilson, 1972).

II.3.4. The Erratics of South Australia

Although Moore (1870) described well-rounded quartz pebbles in Cretaceous soils or clays from Wallumbilla (interpreted as concretions by Jack and Etheridge, 1892), the first large middle Cretaceous erratics were described by Brown in 1885 from just north of Lake Gairdner ("*quartzite water-worn boulders,*" Brown, 1885). Later, Jaquet (1893) described similar boulders from the opal deposits at White Cliffs, New South Wales

(N.S.W.). Additional erratics were discovered during the early geological mapping of South Australia by Brown (1892a, 1892b, 1894, 1895, 1898a, 1898b, 1905), who was the first to suggest a glacial origin for them (a view subsequently expounded by Woolnough and David, 1926). There have since been various explanations for these deposits, with a literature extending to the present day. Erratic bearing deposits and pebbly mudstones are now known from the Great Artesian, Eucla, and Perth Basins. Their distribution is summarized in Figure II.9.

The erratics are known principally from the southern margins of the Great Artesian Basin in South Australia and northwestern N.S.W. (Figure II.9). It is this area that has the greatest density of clasts, the greatest variety of compositions (though dominated by quartzites and quartz-rich rocks), and the only evidence of faceting and striae (David, 1932; Ward, 1925; Woolnough and David, 1926). The rarity and faintness of the striae suggested to Woolnough and David (1926) that the striated clasts had undergone subsequent water-borne transport. Furthermore, confusion of faceting with fracture surfaces is a possibility. The erratics are generally well-rounded ("water-worn" of earlier authors, Brown, 1894; Woolnough and David, 1926). It is also in this area (between White Cliffs and Dalhousie) that we find the largest erratic clasts of the Mesozoic and early Cenozoic (>1 m in diameter, up to 3 m in diameter, Frakes and Francis, 1988, although most reports seem to be less than 1.5 m in diameter). Jack (1931), noting the diminution in size and quantity from south to north, suggested that this might reflect the transport direction. Glendonites have been reported associated with these erratics (Anderson and Jevons, 1905; David and Taylor, 1905; Jaquet, 1893; Raggatt, 1938).

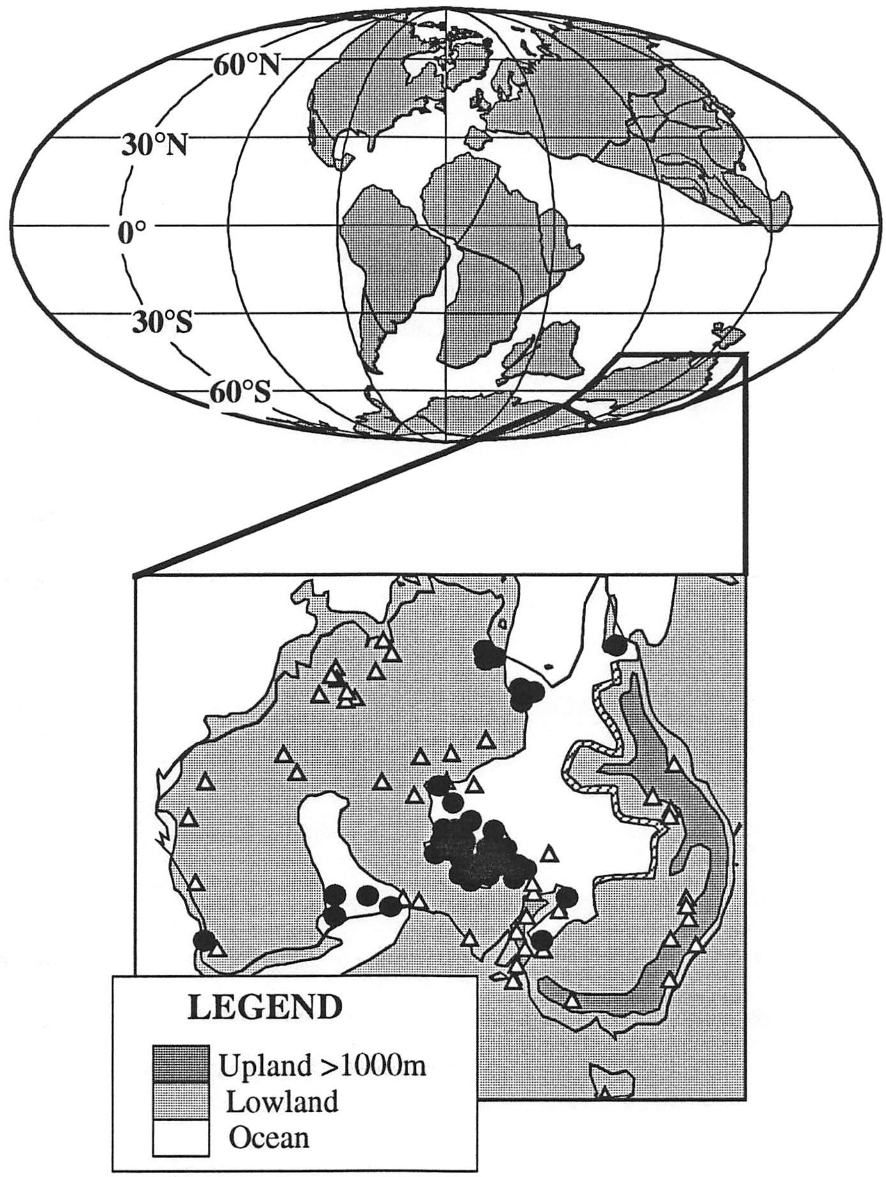


FIGURE II.9. Distribution of Australian middle Cretaceous erratics.

Open triangles = pre-Cretaceous glacial deposits, filled black circles = mid-Cretaceous erratics, diagonally shaded area = pebbly sandstones. See Table II.3 for references. Paleogeography after BMR Paleogeographic Group (1990). Plate reconstruction is for 105 Ma, Albian (Rowley, unpublished data).

Erratics also occur in the Eucla Basin, where the only erratics so far recognized are granite boulders from the Loongana Sandstone (Lowry, 1970) of probable late Neocomian-early Aptian age (Lowry, 1970; Skwarko, 1969). Scattered quartz pebbles are reported from numerous deposits of this age: from the Wyandra Sandstone Member (Aptian) of the Cadni-Owie Formation in the Eromanga Basin (Exon and Senior, 1976), the claystone of the Neocomian-Aptian Leederville Formation (Playford et al., 1976), the Albian-Cenomanian Osbourne Formation in the Perth Basin (Playford et al., 1976), and Unit 6 (Aptian) of the "Coastal Belt suite" of Skwarko (1966; Walpole et al., 1968) in the Northern Territories, which is probably equivalent to the unit where Frakes and Krassay (1992) described erratics. In addition Skwarko (1966) describes isolated pebbles and boulders from the Neocomian marine sandstones of Unit 2 from the same area. Quartz boulders are found in the Rumbalara Shale (Aptian) south of Alice Springs in the Northern Territory. Pebbly sandstones and conglomeratic layers occur along the eastern margin of the Great Artesian Basin associated with the middle Cretaceous transgression, but no demonstrable "erratics" are reported. In the northern Territory the largest clast recorded is only 26 cm in diameter (Frakes and Krassay, 1992).

Sedimentary features associated with the erratics are rarely reported, since most erratics are known from loose surficial specimens. Frakes and Krassay (1992) noted "*penetration structures*" in immediately underlying laminae consistent with a dropstone origin for their erratics, and this has also been reported in South Australia (Frakes and Francis, 1988).

The erratics are commonly associated with fossilized wood (Alley, 1988; David, 1932; Flint et al., 1980; Frakes and Francis, 1988, 1990; Frakes and Krassay, 1992; Krieg, 1986; Mckirdy et al., 1986; Moore and Pitt, 1985; Pittman, 1901; Süssmilch, 1914;

Walkom, 1929; Ward, 1927; Woodard, 1955; Woolnough and David, 1926), with preserved logs up to 21' (6.4 m) long and up to 3.5' (1.1 m) in diameter (Woolnough and David, 1926). Woolnough and David (1926) found a group of logs near Muloowurtina amidst which were numerous small erratics and a large quartzite boulder over 1m long. Fossil wood is also reported from the Perth Basin of Western Australia (Cockbain and Playford, 1972; Playford et al., 1976), Queensland (Day and Tweedale, 1960; Jack and Etheridge, 1892) and Northern Australia (Skwarko, 1966). Concentrations of erratics and wood appear to occur in distinct horizons or boulder beds. These coincide with the basal portions of transgressive units such as the base of the Bulldog Shale and in the Cadni-Owie Formation (Flint et al., 1980). Indeed, the concentration of wood may be diagnostic of transgressive units (Savrda, 1991).

II.3.4.1. Depositional Environment and Provenence

During the middle Cretaceous, a series of major marine transgressions inundated much of Australia (BMR, 1990; Frakes and Rich, 1982). The coals and fluvial deposits of the Late Jurassic and earliest Cretaceous were replaced by marine clays, siltstones and sandstones, which grade into fluvial sandstones and conglomerates around the basin peripheries. Although there were highlands in the east and southeast of the continent associated with the initial rifting between Australia and Antarctica in the south (Otway and Gippsland basins) and between the Lord Howe and Norfolk Rises in the east, most of the land at this time was relatively low lying. There may also have been small islands of Paleozoic rocks, such as Mt. Woods, in parts of the southwestern Great Artesian Basin (Jack, 1915). The land masses were forested, especially in the south, with evidence of upland forests dominated by podocarps (Dettman et al., 1992) during the Aptian. In Victoria a diverse biota including theropod dinosaurs, hypsilphodontids, turtles and birds is known from fluvial and lacustrine deposits of the same period, as are abundant plant remains (Douglas and Williams, 1982; Rich et al., 1988; Rich and Rich, 1989). The Aptian shallow seas appear depauperate in comparison with times before and after; Whitehouse noted the absence of phylloceratid, lytoceratid, hoplitid, or pseudoceratitic ammonites, hibolitid belemnites, reef coral faunas, rudistic lamellibranchs, nerineid gastropods, or large foraminifera (David, 1932; Day, 1969; Scheibnerová, 1971).

Many of the erratics seem to be reworked from earlier diamictite deposits (David, 1907; Ludbrook, 1966; Moore and Pitt, 1985; Parkin, 1956; Woodard, 1955; Woolnough and David, 1926). This may explain both the high density of erratics (especially large erratics) in southern Australia and also the fact that only in this area do Mesozoic erratics show any evidence of striations or facets, the faintness reported for the rare striae reflecting reworking. Flint et al. (1980) suggested that the Devonian quartzite boulders at White Cliffs were originally derived from the Amphitheatre Group near Gobar, N.S.W., and were initially transported from there during the Permian glaciation of the area. Indeed, the clasts of the Cretaceous and those of the Permian tillite are lithologically quite similar. This reworking has probably occurred more than once. Flint et al. (1980) further noted that these clasts are not restricted to the Bulldog Shale but also occur in the underlying Mt. Anna Sandstone, Cadni-Owie Formation and the fluvial Algebuckina Sandstone (nomenclature as given in Flint et al., 1980). Slumping is present in the Bulldog Shale (Flint et al., 1980; Wopfner et al., 1970). As with the erratics of the Chalk, it is probable that such material littered shorelines during this period (Flint et al., 1980; Hawke and Bourke, 1984; Wopfner et al., 1970).

II.3.5. Erratic Emplacement Mechanisms.

II.3.5.1. Non-rafting Mechanisms

II.3.5.1.1. Impact

Oberbeck et al. (1993) and Rampino (1992) have recently drawn attention to the purported resemblance between diamictites of glacial and those of impact origin. While some previously described "glacial" diamictites may indeed be impact ejecta, established glaciations with their repeated glacial cycles, geomorphic landscapes (glacial valleys, nunataks, etc.) and facies transitions, are difficult to reinterpret as impact related, and we find no support for this alternative interpretation. An impact origin for the pebbly mudstones and erratic bearing deposts of the Mesozoic-early Cenozoic is untenable.

II.3.5.1.2. Mass flow

Mass flow deposits form in geological settings where gravitational instabilities exist. They are typified by diamictites (*"terrigenous sedimentary rocks that contain a wide variety of particle sizes*," Flint et al., 1960) and are more or less recognizable depending on the characteristics of the flow regime responsible. Thus the presence of slumps or graded stratification (indicative of plastic and viscous fluid flows respectively) is generally diagnostic, but intermediate flow regimes (fluid enough not to preserve slump structures, but cohesive enough to prevent total separation and differential resettling) can be ambiguous. Such intermediate flow regimes are typified by pebbly mudstones (see (Crowell, 1957; Dott, 1963; Middleton and Hampton, 1973; Stanley, 1969), which are the major facies used in the Mesozoic to infer glaciation (Epshteyn, 1978; Frakes and Francis, 1988).

The superficial resemblence between diamictites of glacial and mass flow origin is problematic, especially in the absence of lateral facies constraints or demonstrable glacial striae. This has led some authors to suggest that many "glacial" deposits may in fact be mass flow deposits and should be re-evaluated (Crowell, 1957; Crowell and Winterer, 1953; Dott, 1961), such as the Squantum "tillite," Massachusetts (Coleman, 1926; Dott Jr, 1961), Ridgeway and Gunnison "tillites," Colorado (Van Houten, 1967), middle Caradocian Cosquer Formation, Brittany, France, (Long, 1991) and the Precambrian Poudingue de Granville of Normandy (Winterer, 1964). Where such deposits are associated with turbidites and slumped units, as in the Campanian-aged Lago Sofia Conglomerate (Cerro Toro Formation) of Chile, this re-interpretation seems justifiable (Cecioni, 1957; Scott, 1966). However, a study of Antarctic depositional systems shows that the two are not mutually exclusive (Kuvaas and Leitchenkov, 1992; Wright et al., 1983). Indeed, Heezen and Hollister (1964) proposed a link between periods of high turbidite deposition and periods of glaciation. Nonetheless, whereas glaciation may lead to increased physical weathering and sediment supply, resulting in high volumes of sediment on the continental shelf and concomitant instability, the reverse need not be true. Without corroborating evidence, mass flow is no more than evidence of gravitational instability and slope failure and not necessarily of glaciation.

The lack of any direct association between slumping and erratics in either South Australia or the Chalk of southern England would suggest that this is not the principle emplacement mechanism in either case, although it may have been responsible for reworking the clasts once they were deposited. In both cases the clasts must be transported into the depositional basin by another process and this can only be through rafting.

II.3.5.2. Rafting Mechanisms

II.3.5.2.1. Ice

Glacio-marine sediments are currently forming in ≈10% of the world's oceans (Drewry, 1986). During the last glaciation, such deposition accounted for far greater volumes of sediment. In the Pleistocene North Atlantic up to 40% of the sediment deposited was transported by ice rafting (Molnia, 1983). A distinction must be made between iceberg rafting (evidence of calving glaciers) and sea or coastal ice rafting (indicative of at least seasonal freezing).

II.3.5.2.1.1. Coastal, sea and river ice.--Coastal, sea and river ice form by in-situ freezing. Although such ice makes no contribution to eustacy, it has been used as evidence of prevailing cold conditions and possible glacial conditions in higher latitudes. It has also been suggested as a rafting agent to explain Cretaceous erratics (Frakes and Francis, 1990; Lyell, 1872).

Observational data suggest that ice growth is a function of the number of successive days with air temperatures below freezing (more realistically taken as -1.8°C, the freezing temperature of "average" sea water, Bilello, 1961; Figure II.10). This initially occurs in shallow or low salinity regions such as coasts or especially river mouths (Markham, 1986), where sediment is most likely to be entrained. Although today perennial sea-ice is essentially limited to the Arctic Ocean and a narrow belt around the Antarctic continent (Gow and Tucker, 1990), seasonal freezing of water bodies does occur elsewhere. Historical records report ice formation in low latitudes during particularly severe cold periods, such as the Baltic (Lindgrén and Neumann, 1982), the Venetian lagoon (Camuffo,

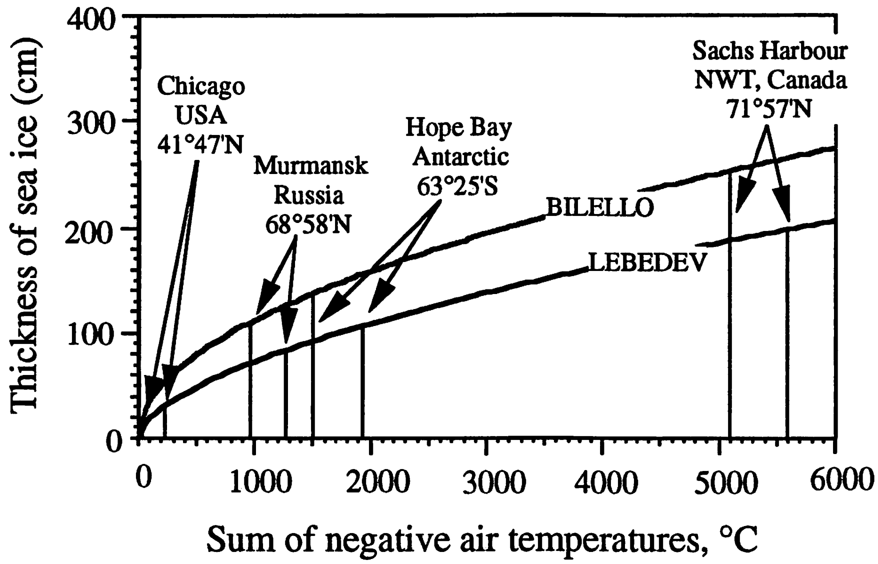


FIGURE II.10. The relationship of ice thickness to its effective growing period.

The effective growing period is approximately represented by the sum of mean daily temperatures for consecutive days with temperatures below either 0°C, as used by Lebedev, or -1.8°C used by Bilello (1961); -1.8°C more closely approximates the freezing temperature of sea-water.

1987), and the Suffolk coast (Lamb, 1985), and in 829AD ice was even reported on the Nile River (p.427, Lamb, 1985). Such low latitude sea and river ice is generally thin reflecting the limited time available for growth. Since deposition of the Chalk occurred at about 45° paleolatitude, rafting by sea- or river- ice would imply that seasonal thermal conditions during the middle-late Cretaceous were similar to, or indeed colder than, those of the present. These conditions are incompatible with fossil evidence.

Although we can estimate the carrying capacity of sea or river ice (Figure II.11), observations show that the limiting factor for rafting of debris is in-situ melting, not

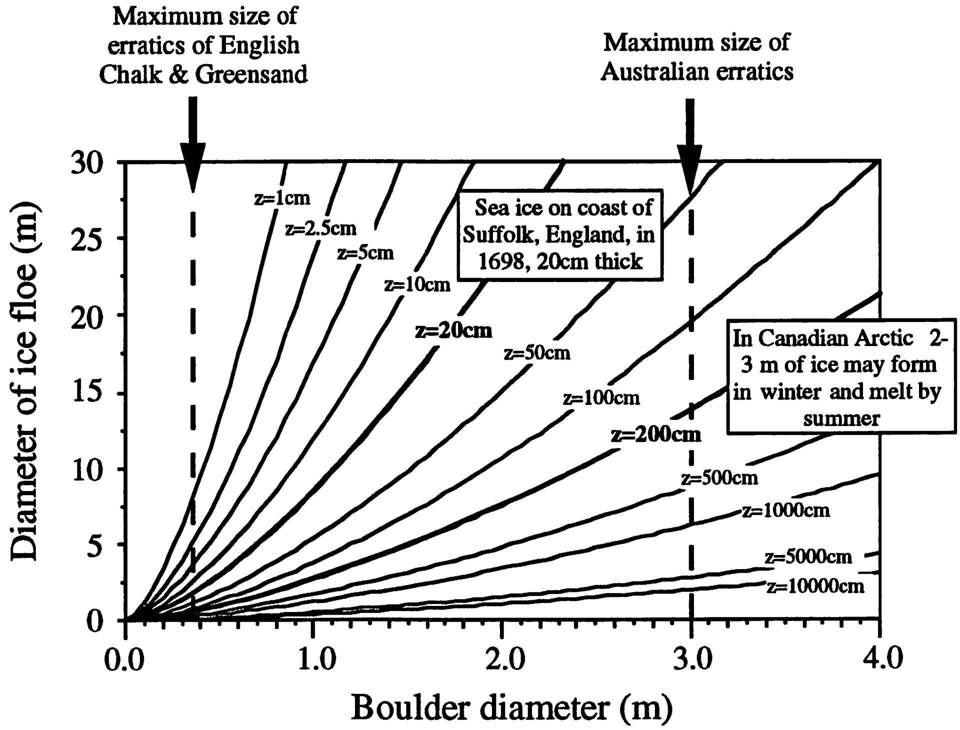


FIGURE II.11. The carrying capacities of ice bodies of different thicknesses (cm) and areas (m).

The area of ice required to carry boulders of specified sizes is represented as a circle with the diameter given on the y-axis. Note that sea-ice loses most of its entrained material in-situ during initial melting.

carrying capacity. Entrainment of coarse material occurs mostly through basal freezing adjacent to the sediment-water interface (Osterkamp and Gosink, 1984). Since this is also the site of early melting (indeed such melting is a prerequisite if the ice is to move away from the shore), most coarse material is lost before significant transport has occurred (Clark and Hanson, 1983; Kempema et al., 1989; Wollenburg et al., 1988). Studies in the Beaufort Sea of Alaska show that in-situ melting accounts for the loss of 90% of entrained sediments before any transport occurs (Drewry, 1986). Similarly, Drake and McCann (1982) found that boulders were also deposited in-situ on Canadian tidal flats during the early stages of breakup and melting. Here, sliding is the dominant mechanism for moving large boulders, a process that is precluded when the water depth is greater than the diameter of the boulder. Quantitatively, only fine material (silt or finer) is transported more than a few 100 meters from the coast, and the sediment content of sea ice is dominated almost exclusively by eolian derived material within a few kilometers offshore (Drewry, 1986). Wollenburg et al. (1988) noted that the sea ice north of Svalbard comprised 40-60% clay and 30-50% silt with a maximum of only ≈10% fine sand. Sharma (1979) found that sea ice sediment from the Bering sea comprised mostly silt (35-80%) and clay (12-60%) with only minor sand (1-8%), while Hoskin and Valenica (1976) noted that ice rafted sediment accumulating on the floor of Yakuta Bay, Alaska, is mainly mud. Thus although the high paleolatitude of South Australia in the Mesozoic is not necessarily incompatible with seasonal freezing (Dettman et al., 1992), the large sizes of the clasts (especially those >1 m) precludes rafting by such ice.

II.3.5.2.1.2. Icebergs.--The size of transported clasts is not a limiting factor for iceberg rafting, although like sea ice, debris in icebergs is dominated by the finer fraction (e.g. Kane cores, Pacific Ocean: coarse pebble fraction comprised <1% of total sediment; silt and clay fraction average 77.2% of all sediment, Molnia, 1983), but the method of entrainment differs. Glacially derived material enters the ice system either through subglacial erosion or from material falling onto the ice surface (supraglacial). Once entrained, material can be moved throughout the glacier by internal deformation. Observations have shown that Arctic derived icebergs typically contain much greater amounts of debris throughout the iceberg than do those from Antarctica (Anderson et al., 1980), although quantitatively the basal layers are the most important source of rafted material (Dowdeswell and Dowdeswell, 1989). Arctic glaciers are generally small valley types with adjacent mountains supplying morainic material onto the glacial surface. Such supraglacial debris is generally absent in Antarctica, where the large ice sheets have largely overridden the valley sides. This difference is reflected in the character of glacial debris. Material from basal levels has undergone mechanical breakdown and sorting during transit (Anderson et al., 1980; Domack et al., 1980). It is typically subrounded to rounded (Dowdeswell and Dowdeswell, 1989; Gottler and Powell, 1989) and is dominantly fine grained close to the base (Drewry, 1986). Supraglacial material which has not experienced communition is typically angular and poorly sorted (Dowdeswell and Dowdeswell, 1989; Gottler and Powell, 1989). This distinction can be used in the geological record for distinguishing between alpine type glacial deposits and those formed from icesheets (Matsch and Ojakangas, 1991). Mesozoic and early Cenozoic erratics (see Table II.3) are dominated by smoothed and rounded clasts, such that if they were transported by icebergs the source of that ice would have been large ice-sheets rather than localized mountain (alpine) glaciers.

Ultimately, melting is the limiting factor in the distribution of ice rafted debris (IRD). Calculations of melting rates suggest that on average 1 m of basal ice is lost every 66 days (Drewry, 1986), although rates range between 0.1 and 100 m per year (Drewry and Cooper, 1981). As a consequence, most iceberg rafted material is lost early, as in Spitsbergen where little IRD makes it out of the fjords. Studies in the fjords of the Antarctic Peninsula show that the percentage of sand decreases exponentially with distance from the ice front (in Lapeyrere Bay the amount of sand in surface sediments drops from ≈80% at the ice front to less than 5% 15 km down fjord, Domack et al., 1989). Furthermore, most of the basal material carried by the inland icesheets in Antarctica is lost by melting close to the grounding line (Drewry and Cooper, 1981) and is not present in the ice shelf or calved icebergs.

Icebergs form by calving from glacial termini. If the entrained material is to be deposited in marine sediments then these termini must be at sea-level and the ELA will be at low elevations which, according to the present glacier distribution (Figure II.5), suggests that land regions poleward are more or less fully glaciated. This is unrealistic for the erratics of the Chalk, but is also difficult to accept for those of South Australia because such glaciers would require that all of contemporary Antarctica be glaciated for which there is no demonstrable evidence. But is iceberg rafting really necessary? The answer is unequivocally no, because there are other viable rafting agents available.

II.3.5.2.2. Animals

Many organisms, including marine mammals, crocodiles, fish and birds, are known to transport pebbles in their stomachs, either as an aid to digestion or as possible ballast (Emery, 1963; see also Table II.4). Emery (1963) reported finding stones in the stomachs of 73 of 170 California sealions studied (Emery and Tschudy, 1941; Fleming, 1951). In the fossil record, "gastroliths" have been found associated with marine reptiles, notably the plesiosaurs (Bearpaw Shale, Montana, Darby and Ojakangas, 1980; Pierre Shale, South Dakota, Welles and Bump, 1949; and Gault Clay, England, Seeley, 1877). A study of the Eocene vertebrates of the Messel Shale of Germany noted that almost all adult specimens of the alligatorid genus *Diplocynodon* contained a number of small pebbles (Schaal and Ziegler, 1992). Described gastroliths are invariably rounded and smoothed, although sizes vary depending on the organism responsible. The sealion gastroliths described by Fleming (1951) range up to 9.4 cm long, while gastroliths found in intimate association with Cretaceous plesiosaur remains range up to 12.8 cm (Welles and Bump, 1949) (see Table II.4). Thus rafting by marine reptiles can explain most of the erratics of the Mesozoic and early Cenozoic, and indeed a group of 107 stones, interpreted as gastroliths, have been found in the Chalk at Rochester (Hawkes, 1951). However, it is

TABLE II.4. RAFTING MECHANISMS

Mechanism	Sizes	Sorting	Shape	Other
Icebergs	clay to boulders[1]	poor[1]	angular (supraglacial), to rounded (subglacial)	facets, striations may be present[1]
Sea/coastal ice	clay to boulders[1], dominantly wind blown material (silt & fine sand)	poor to good[1] (beach derived may be very well sorted depending on source)	angular to rounded[1], depends on source. beach derived material generally rounded[11]	some striations[11], more proximal to source than icebergs [1,6] (i.e. coastal). Organic matter is rare[11], Carbonate content low[11]
River Ice	clay to boulders, dominantly wind blown material	poor to good, depends on source	depends on source, but generally rounded	deposited close to river mouth, associated with other fluvial material, terrigenous input
Mammals	up to 9.4cm long recorded (sea lion)[9]	depends on source	generally sub-well-rounded[2], but depends on source	may occur in groups depending on amount of postmortem reworking; occasionally found with remains of organism; very proximal to source[7]
Reptiles	up to 12cm[2] (pebbles associated with Cretaceous Elasmosaurs are between 1 and 12.8cm, mostly between 2 and 8cm[10])	depends on source	generally rounded	may occur in groups depending on amount of postmortem reworking; occasionally found with remains of organism[10,12,13]
Fish	up to 10cm diameter[2], generally smaller		generally rounded	

TABLE II.4., continued.

Mechanism	Sizes	Sorting	Shape	Other
Trees	up to at least 400cm diameter[3]	depends on source	depends on source (especially whether they are derived from a soil (river - beach bank) or via tree-root dislocation along joints (may be more angular)	widespread[3], clasts generally isolated, but aggregations reported[3]
Seaweed	up to 30cm diameter, depending on size of alga (see discussion)	depends on source	often rounded[2], but depends on source	generally isolated clasts, but some aggregations reported[4], evidence of boring and encrusting organisms may be present[4], the larger seaweeds are restricted to higher latitudes[3]
Birds	up to c.3cm diameter[2], usually <1cm[2]		rounded	
Flotation	pumice (various), sand and silt, coral		depends on source, can be angular (pumice)	pumice can float long distances[2], sand requires very quiet water,
NON - RAFTING MECHANISMS				
Aeolian	upto 200μm for long range transport[8]	very well sorted	well rounded[2], frosting may be apparent	widespread, Chinese sediment found on Hawaii[8]
Mass-Flows	up to kms (olisthostromic clasts)	depends on source	depends on source	associated graded beds, slumps

REFERENCES

1 - Gilbert, 1990; **2** - Emery, 1963; **3** - Emery, 1955; **4** - Emery & Tschudy, 1941; **5** - Church, 1919; **6** - Clark & Hanson, 1983; **7** - Emery, 1941; **8** - Betzer et al., 1988; **9** - Fleming, 1951; **10** - Welles and Bump, 1949; **11** - Lisitzin, 1958; **12** - Darby and Ojakangas, 1980; **13** - Nakaya, 1989.

difficult to explain the 3 m diameter boulder found in the Bulldog Shale of Australia as a gastrolith, and thus alternative mechanisms must be operating.

II.3.5.2.3. Seaweed

During his voyage on H.M.S. Beagle, Darwin noted that certain seaweeds could carry pebbles and boulders far into the ocean (Darwin, 1839; p.239, 1896). Kelp (the common name for any of the members of the brown algae orders Laminarialis and Fucales) are commonly up to 50 m in length (p.151, Lüning, 1990), and can occasionally reach 200 m long (Emery and Tschudy, 1941). The carrying capacity of kelp is a function of its size and the strength of the holdfast (the rootlike structure at the base of the seaweed). Gilbert (1984) noted that of the 109 stones he studied which had attached *Fucus* specimens, ≈60% were in the size range 3 - 6.5 cm diameter, with the largest 9 cm in diameter and weighing 1.2 kg. It is not uncommon for one pebble to have more than one alga attached and consequently the combined carrying capacity is much greater. Church (1919) reports one case where nine large Laminarians were found with stones weighing over 56 lbs (25.5 kg). Church (1919) further states that *Saccorhiza bulbosa* frequently carries blocks of 50-60 "or even 80 lbs" (23-27, 36 kg). Darwin's (1896) description of some of the attached stones as so heavy "*that when drawn to the surface, they could scarcely be lifted into a boat by one person*" is thus not improbable. Once seaweed reaches neutral or positive buoyancy (and even as it sinks) it is at the whim of ocean currents (Gilbert, 1984). Masses of seaweed have been observed far from land: for example, Emery and Tschudy (1941) report floating kelp 400 km from the nearest land; Menard (1953) reported a mass of floating seaweed (*Nereocystis*) ≈30' (c.10 m) long just east of Gilbert Seamount, or ≈1000 km from the nearest land (Alaska) and 480 km from the nearest shallow water where it could have grown (Menard, 1953). One plant in this mass held a fine-grained calcareous sandstone cobble which measured 8 x 5 x 4 cm (Menard, 1953).

Material is released either by sinking of the kelp, due to changes in water density or saturation of cells and buoyancy organs, or by the severance of the holdfast from the stipe (stem) due to organic activity (consumption by invertebrates, fish etc., Emery, 1963). Edgar (1987) noted that the holdfasts of the kelp *Macrocystis pyrifera* observed off Tasmania were rapidly destroyed by the abundant boring isopods, *Phycolimnoria spp.* Upon reaching the sediment surface the organic matter may be destroyed very quickly through biodegradation or consumption by benthic invertebrates (Gilbert, 1984). Consumption by echinoids and gastropods is especially significant, and as a consequence, evidence about the nature of the transporting medium may be lost. Seaweed also transports numerous organisms including echinoderms, sponges and worms, bryozoa, foraminifera, worm tubes, and calcareous red algae (Edgar, 1987; Emery and Tschudy, 1941).

Unfortunately the fossil record of seaweed is extremely poor and although, like marine reptiles, rafting by seaweeds could potentially explain all but the very largest erratics of the Mesozoic and early Cenozoic, direct support for this mechanism is presently absent.

II.3.5.2.4. Trees

Tree rafting has often been considered an explanation for erratics in the geologic record (Lyell, 1841), but has generally been dismissed in preference for an ice origin because of purported size restrictions (Bonney, 1872; Double, 1931; Frakes and Francis, 1988; Frakes and Krassay, 1992; Godwin-Austin, 1858; Jeans et al., 1991; Lyell, 1872). However, the carrying capacity of trees depends entirely on their size and bulk density; the larger the tree, the larger the boulder that can be carried. We can place some constraints on the carrying capacity of wood using Archimedes' principle to assess the length of tree necessary (given wood density and trunk diameter) to carry a specified boulder (the density of granite, $2.7 gcm^{-3}$, is used). The derived equation used in this calculation is:

$$L_t = ((D_b^3 / 6) * ((\rho_b - \rho_w) / (\rho_w - \rho_t))) / R_t^2 \qquad (3)$$

where L_t is the length of tree required (cm), R_t is the radius of the tree (cm), D_b is the diameter of the boulder (cm), and ρ_b, ρ_w andρ_t are the densites of the boulder, water, and tree respectively (g/cm^3). The results for various tree radii are shown in Figure II.12. A similar calculation was made by Liu and Gastaldo (1992) to assess the carrying capacity of trees in the Carboniferous upper Pottsville Formation of Alabama, although they assumed that log-rafts rather than individual logs were required. Curves represent a range of density values that take into account taxonomic differences as well as different densities resulting from dry wood and water saturated wood.

Historical evidence of tree rafting illustrates the large distances and carrying capacities possible. Darwin (1839) described boulders on the Keeling Islands transported by trees. More recently a 3 m (10') long *Nothofagus* log with embedded pebbles was discovered on a beach in SW Tasmania and which was probably derived from South America, some 16,000 km away (Barber et al., 1959). Emery (1955) reported a sighting of a 35 m tree in the Pacific which held a 3-4 m boulder; the sighting was 240 km from land and the log well encrusted with boring organisms and barnacles. Although trees ultimately sink through water saturation of their tissues and boring (Emery, 1963), the distances described above show that this process is not necessarily rapid. Indeed, Pratt (1970) proposed that the erratics dredged from the Blake Plateau were rafted in trees from the Guyana Shield (the largest erratic in this case was 15 x 25 x 20 cm). In addition to isolated

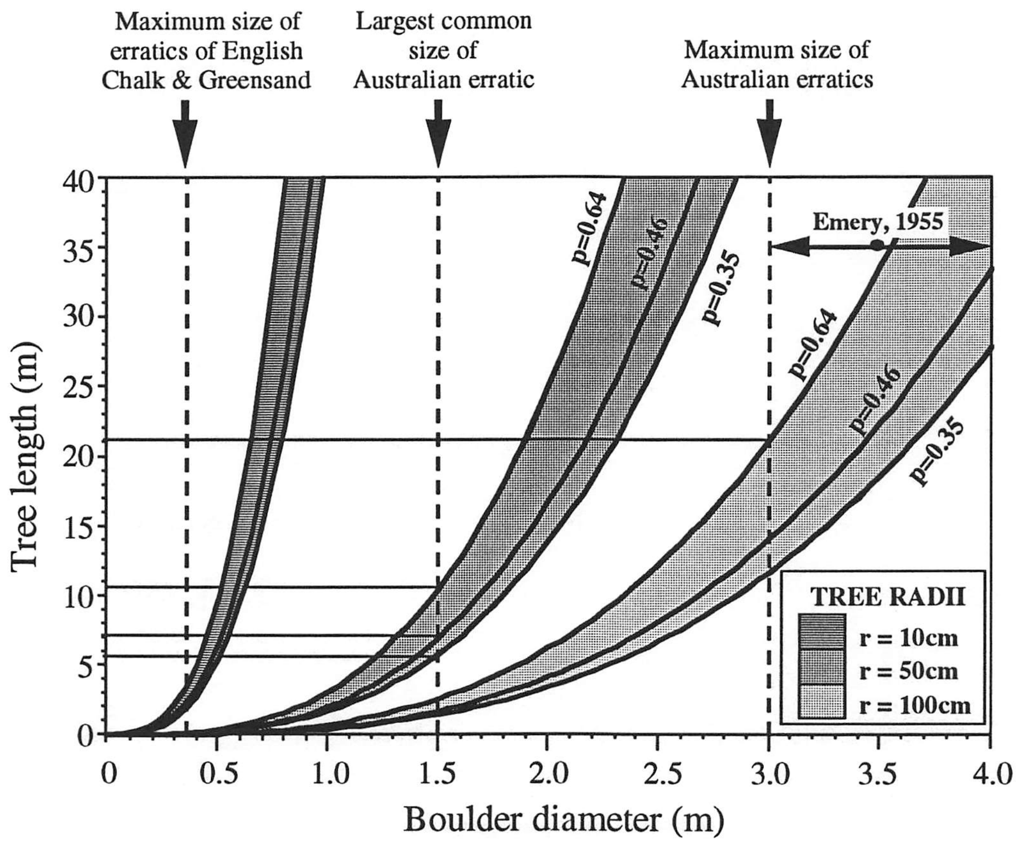


FIGURE II.12. Graph showing the tree size required to carry boulders of specified size.

The range of tree densities, p, reflect the range of densities in living gymnosperms given moisture contents of 12% (Niklas, 1992). Saturation of the wood during transit will increase the density. The position on the diagram of the 35m tree with a 3-4m diameter boulder mentioned by Emery (1955) is indicated by the arrowed line (top right).

boulders or pebbles, sediment can also be caught in the roots of trees and transported (Markwick, personal observation).

From Figure II.12 it is clear that all of the erratics listed in Table II.3, including the largest clasts of South Australia, could be transported by trees. Since fossil wood is extremely common in Mesozoic and early Cenozoic erratic-bearing and associated deposits, particularly those of Australia (as discussed earlier), it is surprising that it has been so readily dismissed. In part, this dismissal of tree-rafting results from the observation that the preserved wood fragments are incapable of carrying the associated erratics (Frakes and Francis, 1988; Godwin-Austin, 1858; Woolnough and David, 1926). This is untrue, since the preserved "fragments" are undoubtedly smaller than the volumes of contemporary wood available because of taphonomic factors. The rarity of wood and floral material in the Chalk, for instance, despite a good vertebrate and invertebrate record, is almost certainly more a consequence of the alkaline conditions prevailing in a carbonate-rich environment than an indicator of the absence of vegetation on surrounding landmasses. Wood is certainly known from flints preserved in the Chalk (Cayeux, 1897; Dines et al., 1933; Dixon, 1850; Jukes-Browne and Hill, 1887; Lyell, 1841; Phillips, 1865; Stebbing, 1897), which suggests that it was present on adjacent land. Interestingly, wood from the Chalk is commonly water-worn and bored (Dixon, 1850; Mantell, 1833) consistent with rafting in the oceans. *Teredo*-like borings are also common in middle Cretaceous fossil wood from South Australia (David, 1932).

Tree-rafting consequently provides an excellent potential explanation for the erratics of the Mesozoic and early Cenozoic, supplemented with other organic rafters. Consequently, in the absence of any other compelling evidence for ice-rafting, there is no necessity for continental glaciation in the Mesozoic or early Cenozoic to explain such deposits. Thus, we are left with only two lines of evidence: the 3rd-order curves of Haq et al. (1987) or their equivalents, which we have already shown are problematic, and the geochemical evidence for ice-sheets.

II.3.6. Geochemical Evidence: Oxygen Isotopes as an Indicator of Sequestered Water Volumes.

It has been known since the 1950's that gradients in the concentration of ^{18}O observed in the surface waters of the world's oceans are primarily controlled by evaporation, hence the apparent relationship with changes in salinity. The continuous loss of ^{16}O-rich freshwater to ice regions leads to a relative enrichment of the oceans in ^{18}O (Epstein and Mayeda, 1953). Since the amount of enrichment is proportional to the volume of water sequestered, the record of ^{18}O in seawater ($^{18}O_w$) through time should also provide a record of glacial volume changes, and thus a potential means of testing the validity of proposed glacio-eustatic sea-level curves (Matthews, 1984; Rowley and Markwick, 1992). However, the direct measurement of ancient seawater is not possible. Instead $^{18}O_w$ values are determined by analysizing the isotopic composition of carbonate preserved in fossil shells ($^{18}O_c$). Unfortunately this composition reflects not only $^{18}O_w$, but also the temperature at which the carbonate was precipitated (Epstein et al., 1951; Epstein et al., 1953; Urey et al., 1951) such that:

$$T(°C)=16.9 - 4.38\ (\partial^{18}O_c - \partial^{18}O_w) + 0.1\ (\partial^{18}O_c - \partial^{18}O_w)^2 \qquad (2)$$

where T(°C) is the temperature and $\partial^{18}O_c$ and $\partial^{18}O_w$ are respectively the isotopic composition of calcite and sea-water relative to some standard. Thus to calculate $\partial^{18}O_w$, T(°C) must be specified. Matthews and co-workers attempted this by assuming that surface temperatures in the tropics have remained essentially constant at ≈28°C and only analyzing tropical plankton (Matthews, 1984; Matthews and Poore, 1980; Prentice and Matthews, 1988). A consequence of this approach is that for ice-free conditions ($\partial^{18}O_w$ = -1.2‰, the result of melting the present ice sheets, Shackleton and Kennett, 1975) values of $\partial^{18}O_c$ for planktonic tropical forms must be ≤ -3.0‰ (relative to PDB). This implies that ice-sheets

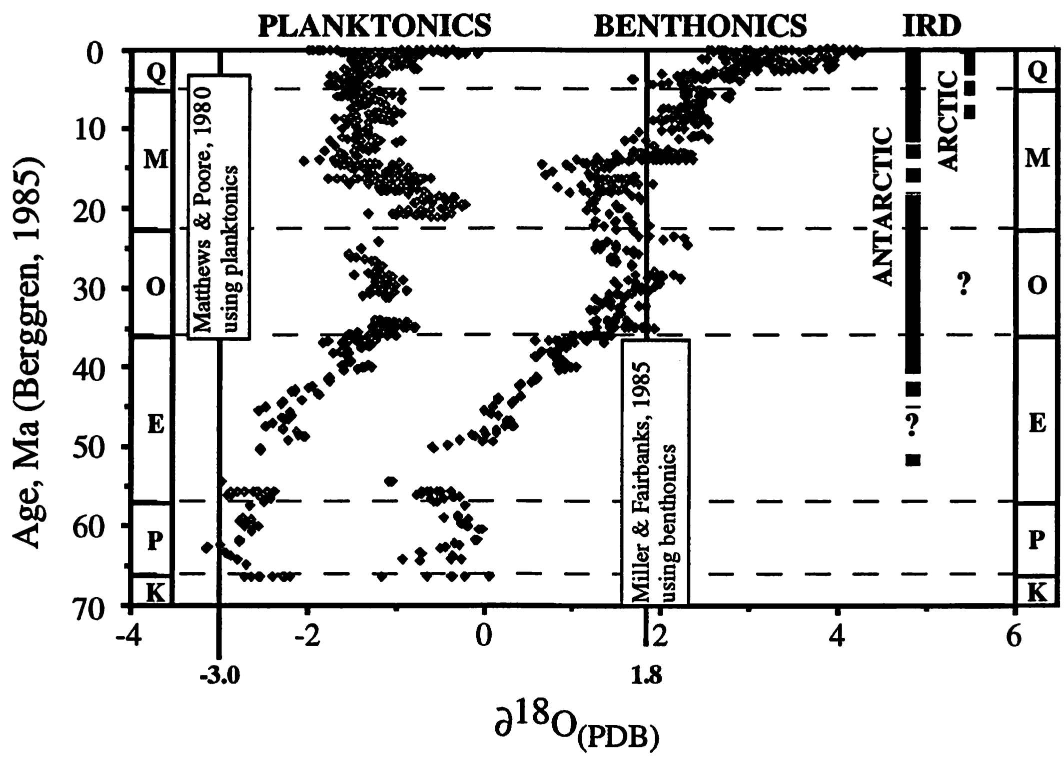


FIGURE II.13. The benthonic and planktonic oxygen isotope curves (after Prentice and Matthews, 1988).

occupied high latitudes at least back into the Maastrichtian and possibly throughout the Cretaceous (Matthews and Poore, 1980; see Figure II.13). This conclusion differs from that of Miller et al., (1987) in which ice-free conditions are implied prior to the Oligocene using benthic rather than planktonic values of $\partial^{18}O_c$ (Figure II.13). This latter method assumes that high-latitude glacial conditions exist when T(°C) for bottom waters is ≤ 2°C, since these temperatures imply that the Antarctic margins were frigid enough for ice-sheet development. This translates to a $\partial^{18}O_c$ of 1.8‰ (PDB), the value for which T (°C) is ≤ 2°C given that an ice-free world has an average $\partial^{18}O_w$ = -1.2‰. Since recent sedimentological data from ODP Leg 119 in the Southern Ocean records evidence of ice-rafting back to the Middle Eocene (Ehrmann and Mackensen, 1992), it is clear that the benthic approach is problematic, although the size of the Late Eocene ice-sheet was probably small (Mackensen and Ehrmann, 1992). Indeed the assumption that bottom temperatures of > 2°C mean ice-free conditions neglects the possiblity that deep circulation during past times was driven by Warm Saline Bottom Water (WSBW) derived in low-latitudes (Brass et al., 1982; Chamberlin, 1906). In such a scenario bottom temperatures do not reflect ambient temperatures in high latitudes and thus make no prediction as to possible contemporary ice-sheets. However, this does not necessarily vindicate the tropical planktonic model since as we have already stated there is no unequivocal glacial evidence prior to the about the Middle Eocene.

That the magnitude of the $\partial^{18}O_w$ shift should be proportional to the volume of water sequestered led Emiliani and Shackleton (1974) to attempt to constrain the isotopic signature of a specified glacio-eustatic change. Using isotopic compositions representative of subtropical sea-surface water and Antarctic snow, they placed bounds of +0.4‰ and +1.6‰, respectively, on the ^{18}O shifts resulting from the extraction of a water volume equivalent to a 120 m drawdown of the oceans (equivalent to sequestering 4.3×10^7 km^3, the volume of water represented by additional ice at the last glacial maximum, LGM; a value of +1.1‰ was used by Emiliani and Shackleton, 1974, as being the most representative for the transition between LGM and present). These bounds correspond to isotopic shifts per volume of water sequestered of 9.30×10^{-9} ‰/km^3 and 3.72×10^{-8} ‰/km^3, which were used by Rowley and Markwick (1992) to retrodict the isotopic signal resulting from the sea-level changes shown by the 3rd order eustatic curve of Haq et al. (1987). If the 3rd order curves are real then the retrodicted signal should be reflected in the actual isotopic record. However, an analysis of the Paleogene reveals no correspondence in either timing or magnitude between the two (Rowley and Markwick, 1992). Although the differences in magnitudes may be partly explained by non-linear fractionation of oxygen during ice build-up (Mix and Ruddiman, 1984), leading Miller et al., 1987, to suggest a value of ≈1.51 x 10^{-8} ‰/km^3 during the inital phase of ice-cap development), this does not explain the disparity in the timing of shifts, and Rowley and Markwick (1992) concluded that such disagreement strongly indicates that the 3rd order eustatic curve of Haq et al. (1987) is invalid.

Comparison between the predicted isotopic curve of Rowley and Markwick (1992) and observed $\partial^{18}O$ shifts is unfortunately mainly restricted to the Cenozoic, for which oxygen isotope data are rapidly accumulating (particularly for Oligocene and younger times through the Ocean Drilling Project, ODP). Reliable data for the Mesozoic remain sparse, in part due to less sampling but also because of problems in determining potential "vital effects" for extinct faunas and increasing diagenetic effects in older rocks. Some data do, however, exist. Barrera et al. (1987) report oxygen isotope data for the Campanian through early Paleocene of Seymour Island, Antarctica, that suggest that shelf waters maintained temperatures between approximately 4.5° to 10.5°C throughout this interval at paleolatitudes between ≈58°S to ≈63°S. This supports interpretations that high latitudes remained ice free. But more significant is the recognition of rapid, high amplitude isotopic shifts in the Cretaceous chalks of Europe that have been interpreted by some workers as reflecting glacio-eustacy (Jeans et al., 1991). Studies by Jeans et al. (1991) and Lamolda et al. (1994) show the greatest isotopic fluctuation occurs in the upper Cenomanian Plenus Marls and lowermost Sussex White Chalk (also known as the Melbourne Rock; Figure II.14), a unit of alternating marls and chalks that coincides with the widely recognized Cenomanian-Turonian Boundary Event (CTBE) and the Cenomanian-Turonian Oceanic

Anoxic Event (OAE; Schlanger et al., 1987). The nature of these oscillations is therefore of great interest. Superimposed on the overall $\partial^{18}O$ shift at this time are fairly systematic isotopic differences between the interbedded marls and chalks which both Jeans et al. (1991) and Lamolda et al. (1994) interpret as deposition of the marls in colder water than the chalks. Similar differences have been reported from the middle Cenomanian chalks of Kent (Ditchfield and Marshall, 1989; mean marl-chalk $\partial^{10}O$ variations of ≈0.4‰). If the overall shift does reflect the development of high-latitude ice-sheets, as Jeans et al. (1991) suggest, then these short-order fluctuations may be interpreted as glacial-interglacial oscillations. Figure II.14 shows the $\partial^{18}0$ data of Lamolda et al. (1994) and estimates of the implied volumes of sequestered water and sea level changes, assuming that the changes in $\partial^{18}O$ are entirely due to ice volume changes. In order to obtain these we assumed that the ice-free conditions are represented by the data at -4‰ and that all deviations are produced by ice formation. We used a linear shift of 3.72×10^{-8}‰/km^3 of sequestered water, equivalent to that suggested by Emiliani and Shackelton (1974) for the Last Glacial Maximum; an average Cenomanian and Turonian elevation of sea level relative to today of 250 m; modern hypsometry from Cogley (1985); a corrected least squares linear regression of elevation and area of $A_{-200 \text{ to } 200m}$ = 106045*E, and 30393700 km^2 for the area between +200 m and +299 m. Even employing a large fractionation effect per volume the Lamolda et al. (1994) data interpreted in this way would imply rather extreme sea level fluctuations of almost 200 m on quite short time scales. We use stratigraphic thickness following Lamolda et al. (1994) rather than time, but the entire interval portrayed may represent no more than ≈1 m.y. Thus the implied fluctuations would be of the same order as those during the Pleistocene. It is worth noting that almost all of the data are more negative than -2.7‰, and even the lightest samples have values -2.0‰. Assuming an ice-free value of $\partial^{18}0_w$ = -1.2‰ and no "vital effects," these values imply sea-surface temperatures of >20-23°C. If significant ice is inferred then the implied sea surface

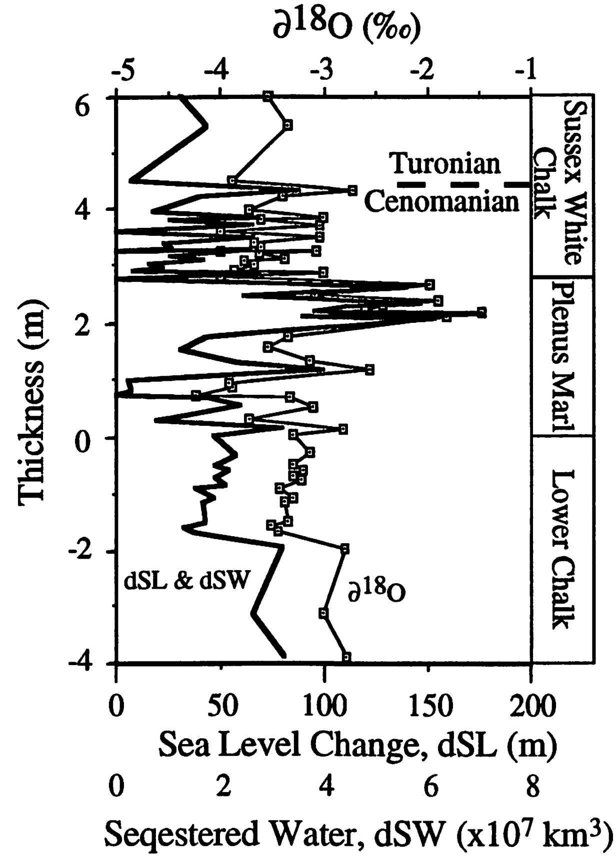


FIGURE II.14. The oxygen isotope curve for the Plenus Marl of southern England (after Lamolda et al., 1994) with the sea-level change and volume of sequestered water implied by these values.

temperatures must increase (>25-28°C). Irrespectively, these implied sea surface temperatures are rather warm for ice-rafting to be considered a viable mechanism for the erratics of the Chalk.

Despite this apparent support for Cretaceous glaciation during Plenus Marl time, there are a number of problems to which we must draw attention. Haq et al. (1987) inferred an average sea-level change of only 25 m for the interval corresponding to the Plenus Marl, far less than that implied by the isotopic fluctuations. The effects of diagenetic alteration, recognized in the Plenus Marl (Jeans et al., 1991), are uncertain, but compensation for more negative diagenetic values may easily diminish the magnitudes of the oscillations and thus any glacial volumes. The relationship between marl deposition, interpreted by Jeans et al. (1991) as representing regressions, and positive $\partial^{18}O$ shifts, that together seem to support a glacial origin, may be of local rather than global origin; indeed the opposite relation was found by de Boar (1982) in the Aptian-Albian marls and chalks near Moria, Italy. The erratics that Jeans et al. (1991) use to corroborate possible Plenus Marl glaciation also occur above and below this unit when there are no isotopic shifts that would seem to require large magnitude changes in either temperature or sequestered water volumes. Lastly, we re-emphasize that shifts of 2 to 3 ‰ are very large indeed and would require Pleistocene or larger volumes of water to be sequestered as ice, for which we have no other record.

Nonetheless, such rapid, high amplitude fluctuations in the ^{18}O record during purportedly non-glacial times are problematic. Although their magnitudes may in part reflect difficulties in isolating the various components of the isotope signal, the coincidence of such sudden purtubations with an OAE suggests to us a possible causal link for both. During purportedly non-glacial periods such as the Cretaceous various workers have suggested that deep-water formation will occur in low latitude epeiric seas rather than at high latitudes as at present (Brass et al., 1982; Chamberlin, 1906). Railsback (1990) has modeled the effect of WSBW formation on the oxygen isotope signal of the worlds oceans and finds that preferential storage of ^{18}O in WSBW could result in a depletion in ^{18}O in contemporary surface waters of ≈0.5 - 1.5‰, similar in magnitude though opposite in sign to that due to the presence of large ice-sheets. Overturning of this salinity stratified system would lead to mixing of ^{18}O-rich WSBW with surface waters and an observed positive shift in $\partial^{18}O$, as seen in the Plenus Marl. Since formation of WSBW is a function of the area of epicontinental seas between ≈10 - 40° N and S (Brass et al., 1982), times of high eustatic sea-level would be conducive to enhanced WSBW formation. Arthur et al. (1987) suggested this might result in increased rates of deep-water upwelling and thus increased productivity and development of organic rich facies. Intermittant overturning of the salinity stratified water column may have resulted from this increased vertical circulation (Southam et al., 1982) or from global cooling due to decreased atmospheric CO_2 concentrations following the burial of large volumes of organic carbon (Arthur et al., 1988; Wilde and Berry, 1982). The rates of WSBW production and oceanic overturn may have been extremely rapid (Brass et al., 1982) and may readily explain the $\partial^{18}O$ oscillations shown in Figure II.14.

From this discussion it should be clear that although oxygen isotopes provide a potential method of examining glacial history, the uncertainities associated with isolating the $\partial^{18}O_w$ signal from other factors remains a problem. We strongly suggest that the ^{18}O signal cannot be interpreted in isolation but must be corroborated with other geological evidence. Nowhere is this better illustrated than in the disparity in ages estimated for the onset of Tertiary glaciation based on isotope data.

II.3.7. Mineralogical Evidence: Glendonites

The mineralogical evidence centers around the interpretation of mineral pseudomorphs collectively known as glendonites. Their sigificance depends on the precursor mineral proposed for the pseudomorph. There have been numerous candidates, including gypsum, $CaSO_4.2H_2O$ (Jaquet, 1893; Pautot and Fontes, 1979); glauberite, $Na_2SO_4.CaCO_3$ (David and Taylor, 1905); thenardite, Na_2SO_4 (Kemper and Schmitz, 1975, 1981); and most recently ikaite, $CaCO_3.6H_2O$ (Shaikh and Shearman, 1987; Shearman and Smith, 1985).

The coincidence of glendonites with glacial erratics in the Permian of Australia (where Dana first descibed such pseudomorphs in the 1840's) implies a link with glacial or at least cold conditions (Brown, 1925; Carr et al., 1989; David and Taylor, 1905; Raggatt, 1938). More recently Kaplan (1977) recognized that glendonites were mainly restricted to high latitudes and often associated with pebbly mudstones, which also suggests a circumstantial link to cold climates. This view has been further expounded by Kemper (1987; Kemper and Schmitz, 1981) and more recently by Frakes and Francis (1990). The suggestion that the precursor might be the mineral ikaite, which is only stable at near 0°C conditions (Pauly, 1963), has consequently received considerable support from these quarters. However, glendonites are also known from low paleo-latitudes, and in particular from the Upper Triassic red bed basins of the eastern United States (Van Houten, 1965). Thus, although an ikaite origin may be appropriate for most glendonite occurrences based on available evidence, it is not likely for all, and a multiple origin for the pseudomorphs seems possible (David and Taylor, 1905). In the case of the Triassic of the eastern United States, the evaporite mineral glauberite seems more suitable; this mineral occurs today in such basins as the Gulf of Kara Bugaz (Kolosov et al., 1974; Sonnenfeld, 1984; Sonnenfeld and Perthuisot, 1989), Death Valley and the Great Salt Lake where it appears to form as a winter precipitate (Warren, 1989).

Low ^{13}C values observed in ikaites, as well as glendonites from the Tertiary of Oregon (Boggs Jr., 1972), imply a bacterial origin for the carbon (Suess et al., 1982), an origin which David proposed for glendonites in 1905 (David and Taylor, 1905). This is consistent with recent experiments on the stability of ikaite by Bischoff et al. (1993) which showed that the formation of significant amounts of ikaite in Ca-dominated cold seawater requires large amounts of dissolved orthophosphate and HCO_3. Both ions are provided by the decomposition of organic matter. We therefore suggest that the distribution of high latitude glendonites during the Mesozoic and early Cenozoic may be as much due to the distribution of organic-rich rocks as to cold conditions. The significance of the association of such organic-rich rocks with transgressive sequences and periods of high sea-level has already been discussed.

II.4. THE PRESERVATION OF TERRESTRIAL GLACIATIONS

Terrestrial glacial deposits have been long claimed to have a poor preservational potential (Bjørlykke, 1985; Croll, 1875; Frakes and Francis, 1990). They are dominantly erosive and deposited material lies largely above local base level where it is subject to reworking and transport, consequently assuming the inherent charcteristics (sorting, shape and size) of the new sedimentological regime, often to the total exclusion of any glacial signature (Croll, 1875). The argument that this preservational bias explains the absence of glacial evidence through the Mesozoic and early Cenozoic (Frakes and Francis, 1988, 1990) is, however, invalid. Bjørlykke (1985) noted that pre-Quaternary glacial deposits are often found in marginal marine or continental rift basins, and suggested that the latter especially would provide favorable environments for preserving glacial sediments. In the Mesozoic and early Cenozoic high latitude rift basins are preserved, such as in Spitsbergen in the Triassic and Jurassic, and more pertinently in southeastern Australia in the middle - late Cretaceous, but they do not contain recognizable glaciogenic sediments. What is more, any purported preservational hiatus as an explanation for the lack of glacial deposits through this 200 million period must also seriously compromise our trust in other areas of the Mesozoic and early Cenzozoic record. Such a hiatus seems unlikely, as these periods are more than well represented in the geological record (Ronov, 1982). In addition, although the details of known glacial "events" are not always clear, especially the further back in time we go, they are nonetheless demonstrable and their effects are entirely consistent with observations of contemporary climate elsewhere on the globe. This would not be true for any purported large-scale glaciation during the Mesozoic- early Cenozoic.

II.5. THE ORIGIN OF MESOZOIC - EARLY CENOZOIC ERRATICS: A DISCUSSION

There is no reason for assigning a common origin for all erratic bearing deposits. Some are clearly due to mass flow (Crowell, 1957; Crowell and Winterer, 1953), such as the Campanian aged Lago Sofia Conglomerate (Scott, 1966), but for most erratics this is not demonstrable and rafting by agents unknown is more appropriate.

Icebergs are the most obvious rafting agents, but their requirement for sea-level terminating glaciers is difficult to reconcile with the established climatic interpretation of the geological record during the Mesozoic and early Cenozoic, especially in the low paleolatitudes of Cretaceous northwest Europe. The faunal (Donn, 1987; Habicht, 1979; Markwick, 1992; Romer, 1961; see Chapter V) and floral data (Dettman et al., 1992; Francis, 1988; Horrell, 1991; Spicer, 1987, 1989; Spicer and Corfield, 1992; Truswell, 1991) clearly imply warmth in latitudes that should, by analogy with the present, have been cooler if an ice-sheet existed. Even if established paleoclimatic interpretations are incorrect, sea-level terminating glaciers in the middle Cretaceous of South Australia implicitly mean a fully glaciated Antarctica with, one would expect, development of terrestrial glacial and glacio-marine sedimentation around its margins. Yet there is no evidence of such sedimentation (Macellari, 1988; Pirrie, 1989). Instead, we know from paleobotanical data that forests covered the coasts of Antarctica from at least the Devonian and persisted well into the Tertiary (Truswell, 1991). Sea-ice rafting, although not incompatible with high latitude paleoclimates per se, even in the Cretaceous (Dettman et al., 1992; Frakes et al., 1992; Frakes and Krassay, 1992; Spicer and Parrish, 1990) cannot account for the size of the clasts preserved, since such material is lost rapidly through in-situ melting before any large-scale transport occurs. Despite this, recent workers have continued to suggest an ice origin for the erratics of the Mesozoic and early Cenozoic (Epshteyn, 1978; Frakes and Francis, 1988, 1990; Frakes and Krassay, 1992; Jeans et al., 1991; Kemper, 1987). As argued in this paper, however, there are alternative explanations.

Organic rafting is a much maligned agent. Frakes and Francis (1988) dismissed it for the erratics of the Bulldog Shale because of the large size of some of the boulders (up to 3 m). This was also the rationale used by Godwin-Austin (1858) and Lyell (1872) for dismissing organic rafting as an explanation of the much smaller erratics of the English Chalk. This reasoning, as we have shown, is invalid. Trees, for instance, are quite capable of transporting large boulders and indeed could transport all of the presently described erratics. The limiting factor is simply the size of the tree (Figure II.12). The close association of erratics and fossil wood in so many erratic-bearing deposits is therefore provocative, such as in the Triassic Lower Beaufort shales of South Africa where Broom (1911) reports of "*...great masses of silicified wood and that a large number of water-worn stones were lying amongst the wood.*" Petrified wood is also occasionally found in the pebble-bearing Aptian Innkjegla Member of the Carolinefjellet Formation of Spitsbergen (Nagy, 1970). The presence of encrusting organisms such as serpulids on many of the erratics of northwestern Europe (Double, 1931; Godwin-Austin, 1858; Kaever, 1974; Sollas and Jukes-Browne, 1873) may also be consistent with rafting by an organic rafter rather than ice. Such organic activity could have occurred after deposition, but this may have been precluded in the Chalk where the "soupy" nature of the substrate (Hancock, 1975) may have resulted in the clasts being enveloped by the sediment; none of the erratics in the Chalk appear to be associated with hardgrounds, with the possible exception of the pebbles in the Totternhoe Stone, Isleham. The Purley locality, in which large cobbles, pebbles and beach sand are found together (Godwin-Austin, 1858) is also significant in this respect, since the association implies the material was gently lowered to the bottom as a group rather than being dispersed during sinking from the sea-surface. In this case, a sinking tree settling to the bottom to rest on the soupy surface seems appropriate. The logs would have been subsequently destroyed by the high alkalinity and organic activity.

What is more, Kaplan's (1977) recognition of the prevalence of pebbly mudstones in high latitudes may reflect not only the potential distribution of ice, but also the location of the largest forests and lowest rates of decay. In times without a definable tropical rainforest, such as the Mesozoic and early Cenozoic (Ziegler et al., 1993; 1987), mid-high latitudes become the most productive with presumably the largest trees. The Cretaceous coals of the North Slope of Alaska represent about one third of the recoverable coals of North America (Sable and Stricker, 1987) and were produced at latitudes above 70°N (Spicer and Parrish, 1990). In the Late Paleocene - Early Eocene fossil forest of Strathcona Fjord, Ellesmere Island (≈78°N paleolatitude), there are tree trunks up to at least 1 m and root stocks up to 5 m in diameter and which may have been up to 40-50 m high (Francis, 1988).

There is evidence, however, that other rafting mechanisms may have been responsible. In the Chalk, the small sizes of the clasts mean that they could all have been potentially transported as gastroliths by marine reptiles. The group of 107 small pebbles found together at Rochester (Hawkes, 1951) and the occurrence of a plesiosaur in the same quarry as erratics at Houghton (Dixon, 1850) are suggestive of this agent. Unfortunately, no bones have been found in direct association with erratics in either the Chalk or the Upper Greensand, although they do co-occur in the Gault Clay (Seeley, 1877). Plesiosaurs are also known from the middle Cretaceous of the Great Artesian Basin of Australia (Jack and Etheridge, 1892; Molnar, 1982) and the Campanian-Maastrichtian of Banks Island (Jutard and Plauchut, 1973). The presence of plesiosaurs in high-latitudes may thus explain the occurrence of chert and quartzite pebbles found floating in both the upper part of the Hue Shale (Campanian, pebbles are <4 cm diameter, Molenaar et al., 1988) and the Pebble Shale unit of northern Alaska (Lower Cretaceous, Molenaar et al., 1988). Troedsson (1924) reported on 4 gastroliths (the largest being c.7.1cm long) found with crocodilian remains from the Danian limestones of Malmö, Sweden (see also Deslongchamps' suggestion that the smoothly rounded stones found in the oolitic strata of Normandy were deposited by crocodiles, Godwin-Austin, 1858).

Alternatively, even though kelp is not known from the Chalk, the size range of the erratics is within that explicable by modern seaweed rafting. Kelp forests today are highly nutrient dependent and commonly occur in mid-latitude upwelling zones (Lüning, 1990). In the Chalk seas bathymetric highs appear to have been the sites of dynamic upwelling. Phosphate nodules and coatings are not uncommon. These areas of high productivity may have been the sites of extensive kelp forests, with large numbers of marine reptiles and fish attracted into the area to feed. Erratics carried by such rafters might then be expected to be concentrated around such highs. Lyell (1841) originally suggested kelp or wood for rafting the erratics of the Chalk. Unfortunately, the preservational potential of kelp is so poor that its significance in the past is difficult to assess.

A further matter of interest is the association between erratic-bearing deposits (Table II.3) and marine transgressions. This is clearly seen in Australia where the Neocomian, Aptian and Albian are all times of regionally significant marine inundations (Frakes and Rich, 1982). In South Australia, boulders and cobbles are first concentrated in the initial transgressive phase of the Cadni-Owie Formation (Alley, 1988) and are most abundant in the basal part of the Bulldog Shale marking the major phase of the Aptian transgression (Flint et al., 1980; see Figure II.15). Indeed, the concentration of rafted material along revinement surfaces during such transgressions may account for many erratic bearing deposits. Large amounts of organic rich material are also buried during these transgressions. In the Cretaceous this is particularly true during the Barremian-Aptian-Albian, the Cenomanian-Turonian and to a lesser extent the Coniacian-Santonian (Jenkyns, 1980). These three periods correspond with Kemper's cold intervals, which were defined on the basis of glendonite occurrences (Kemper, 1987), and also include the erratics of the Chalk and South Australia. Positive $\partial^{18}O$ excursions during at least one of these periods (Cenomanian-Turonian) has been interpreted as further evidence for glaciation during this time (Jeans et al., 1991), but as we have shown such shifts can also be explained by oceanic overturning and mixing of surface waters with ^{18}O-rich WSBW, irrespective of any other effects, such as diagenesis, that might be responsible.

Thus a possible scenario develops with marine transgressions resulting in the erosion of forested land masses, especially in high latitudes, supplying the erratics, complete with organic rafts, to the marine record (Figure II.15). In the highly productive epicontinental seas we may envisage plesiosaurs scouping up such clasts to use as gastroliths while large forests of seaweed occupy the nutrient-rich shallows, occasionally to be ripped up by storms. The expansion of epicontinental seas enlarges the area of the photic zone and enhances productivity as well as WSBW formation. This in turn intensifies

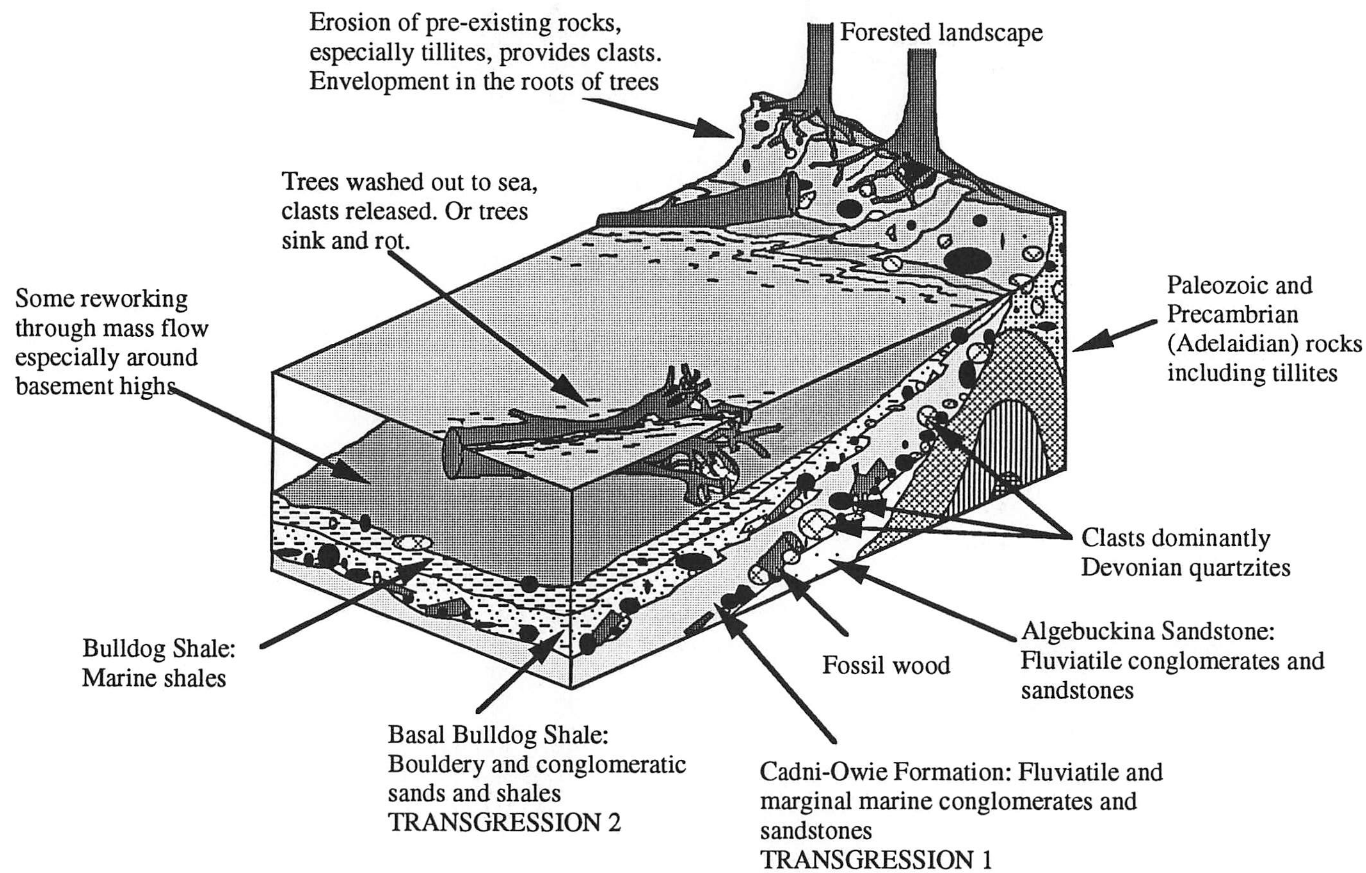


FIGURE II.15. A reconstruction of the possible scenario for formation of erratic bearing beds in South Australia during the middle Cretaceous.

deepwater upwelling and consequently increases productivity of the surface waters, while also in part destabilizing the salinity stratified ocean. Cooling due to the burial of large volumes of organic matter further weakens the system such that intermittent rapid overturning occurs, which appears in the $\partial^{18}O$ isotope record as sudden positive excursions. As for the glendonites, the deposition of organic-rich rocks in the cool bottom waters of high paleo-latitudes would have provided the ideal environment for ikaite formation.

II.5.1. The Causes of Glaciation

If the Mesozoic and early Cenozoic were indeed largely ice free, a possibilty recognized as early as 1836 by Beaumont (1836) and further discussed by Croll (1875), possible causes must be considered. Although a full discussion of this question is beyond the scope of the present paper, we do draw attention to the following: although there are numerous feedbacks that affect global climate, the ultimate driving force for climate change on geologic timescales (>10^6 years) is tectonics (as expressed in changing land-sea distributions and the history of mountain building). The climatic effects of changing land-sea distributions were noted by Lyell, following the ideas of Baron von Humboldt:

> "*...whenever a greater extent of high land is collected in the polar regions, the cold will augment; and the same result will be produced when there shall be more sea between or near the tropics; while, on the contrary, so often as the above conditions are reversed, the heat will be greater.*" (p.115, Lyell, 1830).

As stressed at the beginning of this paper, glacier formation requires that land at high latitudes must be of sufficient elevation to intersect the ELA. The coincidence of periods of glaciations and those of mountain building (using area of deformation as a proxy for mountains) is shown in Figure II.7. The significance of land-sea distribution and orography are supported by recent computer modeling results (Barron, 1981; Barron et al., 1984; Ruddiman and Kutzbach, 1989; Ruddiman et al., 1989). Thus the high global sealevels and low topography of the Mesozoic may have precluded the development of large-scale glaciations.

II.6. CONCLUSIONS

If we are to accept the applicability, and magnitudes, of high frequency eustatic changes as implied by the 3rd order Haq sealevel curves, or their equivalents, then we must also accept their implications: that either an unknown eustatic mechanism existed in the past or there must have been a continuous history of waxing and waning of Antarctic sized ice sheets since at least the late Jurassic (Rowley and Markwick, 1992) (Figure II.2). We have shown here that the purported evidence for such ice sheets is ambiguous at best: that the erratic-bearing deposits can be better explained by organic rafting during times of marine inundation than by icebergs; that other cited evidence, such as glendonites, imply only cold conditions with no explicit need for large terrestrial ice sheets. That cold periods may have occurred during so-called "hot-house" periods (Cretaceous) is not at issue. Global climate is not uniform and we stress again that even during non-glacial periods in the Earth history it seems likely that mountain glaciers capped the highest mountain peaks, especially in high latitudes. But such small glaciers would have had little impact upon eustacy.

Despite the fact that over 97% of Antarctica is presently hidden beneath ice, our ability to recognize glaciations during other intervals in the past and also to document the transition into the present "ice-house world" during the Tertiary, suggests that the lack of evidence for glaciations in the Mesozoic and early Cenozoic is because there are no glaciations to recognize. Since the geological evidence around the margins of Antarctica shows no indication of adjacent ice masses, we must conclude that any ice sheets existed far inland. The farther inland they must be placed, the smaller their potential volume and consequently the smaller their impact on eustacy. Therefore, during non-glacial periods it seems unlikely that any rapid changes in eustatic sea-level could have had magnitudes any greater than 10-20 m. On this scale, differentiation between eustatic changes and intrabasinal relative sea-level effects appears questionable. Global correlation using 3rd order curves must therefore be viewed with caution.