The Spatial and Temporal Geomorphology and Surficial Sedimentology of the Gurra Gurra Crescentic Dunes, Strzelecki Desert, South Australia

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A dissertation submitted in fulfilment of the requirements of the

Degree of Doctor of Philosophy, Department of Geology and Geophysics,

University of Adelaide,

1997

Frontispiece: Autumnal transverse dunes at Gurra Gurra waterhole, Strzelecki Desert, South Australia

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Abstract

The Strzelecki Desert's crescentic dunes of north-eastern South Australia (29° 01' S, 140° 02' E) dominate an approximate 2400 m x 320 m area, originating from the crestal zone of a quartz-sand linear ridge at a site known as Gurra Gurra waterhole. The interaction of linear and crescentic crestal morphologies place the linear dune in the complex class-type, however, in part, the identity of the linear morphology is nebulous. Parasitic erosion of the linear dune-form results in contemporaneous re-deposition of the sediment into crescentic dunes, which move sub-parallel to the longitudinal axis of the underlying linear form.

The intricate natures of dune morphology and surficial sedimentology for the Gurra Gurra crescentic dunes are responses to site specific sand sources in a semi-closed geomorphic system, topographic enhancement and seasonal periodicity of less dominant secondary and tertiary wind directions, strengths and durations, as well as other climatic and vegetative influences. The principal influences of multi-directional winds and the upslope-amplification of shear velocities, cause dune form to be in continual transition between the dynamic end-members of equilibrium and quasi-equilibrium. Dimensional equilibrium is a transient feature at Gurra Gurra waterhole and is not characteristic of this dunescape.

The surficial sedimentology of the crescentic dunes is one of medium-to-fine quartz-rich sands with a unimodal, positively skewed and leptokurtic distribution. Analysis has shown that finer mean grain-sizes correlate with better sorting, a less positive skewness and lower kurtosis. The signatures of the sedimentological distributions across the dunes, and in different seasons, are subtle contrasts that are attributed to either the removal or accumulation of the fine tail of the distributions. Seasonal variation is gradual and transitional between each season, while greatest differences are found between the upper and lower micro-geomorphic positions of the dunes.

Through the integration of qualitative and quantitative process-geomorphology, is has been shown that the Strzelecki-Gurra Gurra dunescape portrays a microcosm of aeolian features, and processes, that are not dissimilar to many in the immense areal deserts of Earth and Mars.

Declaration

This work contains no material which has been accepted for the award of any other degree or diploma in any university or other tertiary institution and, to the best of my knowledge and belief, contains no material previously published or written by another person, except where due reference has been made in the text.

I give consent to this copy of my thesis, when deposited in the University Library, being available for loan and photocopying.

Mark A. Bishop, 15 January 1997

Acknowledgements

It is with my sincerest gratitude that I thank the following colleagues and friends who have rallied to guide, assist and debate the work shown herein. With my deepest appreciation I thank-you one and all (Vic Gostin, Bob Bourman, John Cann, and Teresa Dickinson-Laslet), as without each of you, I would never have overcome the many obstacles that have confronted me over the years.

Special indebtedness must also be given to Max Banks, who so long ago, introduced me to the beauty of the earth sciences.

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1 Introduction to Aeolian Geomorphology

1.1 Introduction

The quantitative study of arid zone geomorphology has proliferated in the past fifty years. Since Bagnold's (1941) benchmark work *The Physics of Blown Sand and Desert Dunes,* a torrent of edited review volumes, *A Study of Global Sand Seas* (McKee, 1979a); *Arid Zone Geomorphology* (Thomas, 1989); *Aeolian Sand and Sand Dunes* (Pye & Tsoar, 1990); *Desert Geomorphology* (Cooke, Warren and Goudie, 1993); *The Dynamics and Environmental Context of Aeolian Sedimentary Systems* (Pye, 1993), *Geomorphology of Desert Environments* (Abrahams & Parsons, 1994), *Geomorphology of Desert Dunes* (Lancaster, 1995) to name only a few, exemplifies the many directions aeolian geomorphology has followed in this time. Nonetheless, a fundamental understanding is still lacking about the origin and evolution of many dune morphologies and regional sand seas. For example, hypotheses abound concerning the origin and evolution of the linear dune (also termed the longitudinal dune, sand ridge and seif) and are evidence for the uncertainty about this common, yet morphologically diverse landform. Similarly, the genetic transition and morphological variation of crescentic forms (sand mound, barchan, barchanoid and transverse dunes) are poorly understood. Hence although both vehicular access and satellite remote sensing have made available a plethora of terrestrial localities and numerous extra-terrestrial dunescapes, elementary uncertainties concerning dune origin and evolution still persist.

Breed (1979) commented that fields of crescentic dunes occur in nearly all terrestrial sand seas, a notable exception being Australia, although "it is now suspected that local examples of crescentic dunes do exist amongst the predominantly linear dune forms of this continent." Since the early 1980s, the existence of crescentic dunes in Australian deserts has been documented with notable fields at Gurra Gurra waterhole in the Strzelecki Desert of north-eastern South Australia and along the Finke River in the Simpson Desert of the southern Northern Territory (Wasson, 1983a, 1983b; Mabbut and Wooding, 1983;

Wopfner and Twidale, 1988). However, research on the desert dunes of Australia, regardless of type, demonstrates an absence of spatial and temporal comparisons of morphology and sedimentology, as well as deficiencies in the subsequent integration of these parameters into fundamental interpretations of dune genesis. Despite these inadequacies in the Australian geomorphic context, a plethora of global discussion exists concerning differences in form (Bagnold, 1941; Madigan, 1946; Finkel, 1959, 1961; Long and Sharp, 1964; Norris, 1966; Hastenrath, 1967; Tsoar, 1974, 1978, 1984, 1989; Lancaster, 1978, 1980, 1981, 1982a, 1982b, 1983a, 1983b, 1989, 1994; McKee, 1976, 1979b; Kar, 1987, 1990), the variation in surficial sedimentology between interdune and crest (Bagnold, 1941; Alimen, 1953; McKee and Tibbits, 1964; Sharp, 1966; Glennie, 1970; Folk, 1971b; Lancaster, 1981a, 1982b, 1987) and the dynamic processes involved in the sculpturing of dunescapes (Hastenrath, 1967; Howard *et. al.,* 1978; Sneh and Weissbrod, 1983; Vincent, 1984, Watson, 1986, Livingstone, 1987b; Kocurek *et al.,* 1992). Nonetheless, few of these studies have been designed to evaluate periodic temporal variation of form, surficial sedimentology and process. Cooke *et al.,* (1993, p.341) eloquently voice the current state of research concerning this issue, "..very few of the very large numbers of studies of sorting on dunes have explicitly recognised such complexity, and very few have employed sampling procedures fine enough to discriminate between the various processes. What is worse, many if not most of them [studies] have little or no explicit model of the processes involved. The result is a great body of data that is contradictory and at worst, nearly worthless." Further to this, the understanding of aeolian sedimentological processes would be paramount in the identification and geological comprehension of remnant aeolian environments in the stratigraphic record. Similarly, such understanding would be required in discerning aeolian deposits as resource potentials, as well as the 'natural hazard' they often pose in human occupied territories.

Dune systems are components of a continuously dynamic and changing arid zone landscape that often only reside in a quasi-equilibrated state. Gross change is inevitable and although this may take millennia for mega-forms, this is not the rule for meso- and micro-aeolian landforms, where gross change can be implemented in months, days or even hours. According to Lancaster (1984, 1989) and Warren (1984), the genesis, evolution and interactive mechanism of morphology and dynamics are the least understood parameters in aeolian geomorphology. It is the purpose of this dissertation to assist in the understanding and recognition of the interactive processes in the systematic analysis of the surficial sedimentology and morphology of crescentic dunes. Only by repetitive and sequential examination can transformations be observed for both long term evolution and shorter time scale periodicity within well defined geological and meteorological time frames. In this manner, the diverse and often contrary global information concerning desert landscapes and landforms, will become more unified and predictive through the development of multiple working hypotheses and process-response models that discern the origin and evolutionary cycle of like features, an objective that is achieved in this thesis for the crescentic dunes of Gurra Gurra.

Both depositional (dune) and erosional (yardang) aeolian landforms are common place in terrestrial arid zones, although the terrestrial environment is not the only setting in which they are found. Efforts to understand aeolian processes, desert landforms and landscapes, are not restricted to the terrestrial environment, but have led to investigations of our planetary neighbours. Extra-terrestrial reconnaissance by the spacecraft Mariner and Viking in the period 1965 - 1976, identified Mars to have a landscape richly sculptured by the wind, revealing an abundance of crescentic dune-forms, but an absence of linear morphologies compared to Earth. Subsequently, surveys by Voyager in 1989, imaged aeolian activity on Neptune's largest satellite Triton (Croft, 1990), while the radar mapping of Venus (by the probe Magellan 1990-94), has shown evidence of Venusian dunescapes (Greeley *et al.,* 1991). Undeniably, the union of comparative planetary geomorphology and the use of terrestrial analogues will define a far greater understanding of wind as a geological process, and as used in this dissertation, aid in further understanding the terrestrial aeolian environment. Through the use of terrestrial analogy the origin and evolution of aeolian landforms and processes will be better understood for planetary surfaces enshrouded with an atmosphere.

1.2 Regional Geo-Setting of the Strzelecki and Simpson Deserts

The Australian dunefields dominate an area of some 1,311,000 km^2 (*ca.* 20% of the continent's area) (Ash and Wasson, 1983). The Strzelecki-Simpson dunefields occupy two separate topographic depressions of Cainozoic age, which have adopted their form since the Eocene (Wasson, 1983b). Kopperamanna Gap delineates a junction point between these sedimentary basins where Cooper Creek flows between the Gason Range Anticline (Fig. 1.1). The two dunefields are surrounded by low hills and plains on which both Cretaceous and Tertiary lithologies outcrop. Localised outcrops of Tertiary strata occurs within interdune corridors. However, the majority of the dunefields are aeolian, lacustrine and fluvial-alluvial sediments of Quaternary age. These sediments have originated from the erosion of the area by tectonic influences in the Late Tertiary and the input from large multi-thread streams extending from Queensland and terminating in the deserts (Wasson, 1983b). The relief of the northern drainage pattern and desert basins is only some 20 m per 100 km and develops a multi-thread nature typical of drainage in the Channel Country of Queensland and the Simpson-Strzelecki Deserts. The perimeter of the dunefields shows Early-Mid Tertiary outcrop resulting from uplift, whereby the dunefields occupy morpho-tectonic basins (Wasson, 1983b). Both lie within the Cainozoic Birdsville structural basin (Hills, 1963; Wasson, 1983b).

1.3 Cainozoic Stratigraphy

Jurassic to mid-Cretaceous marine sediments (Winton Formation) of the Eromanga basin underly and outcrop on the perimeter of the Strzelecki-Simpson dunefields. The Cretaceous Winton Formation is overlain by arenaceous fluvial sediments of the Eyre Formation of Palaeocene-Eocene age (Wasson 1983a, !983b). Wopfner (1974) suggests that epeirogenic movements assisted in deposition of the Eyre Formation in the

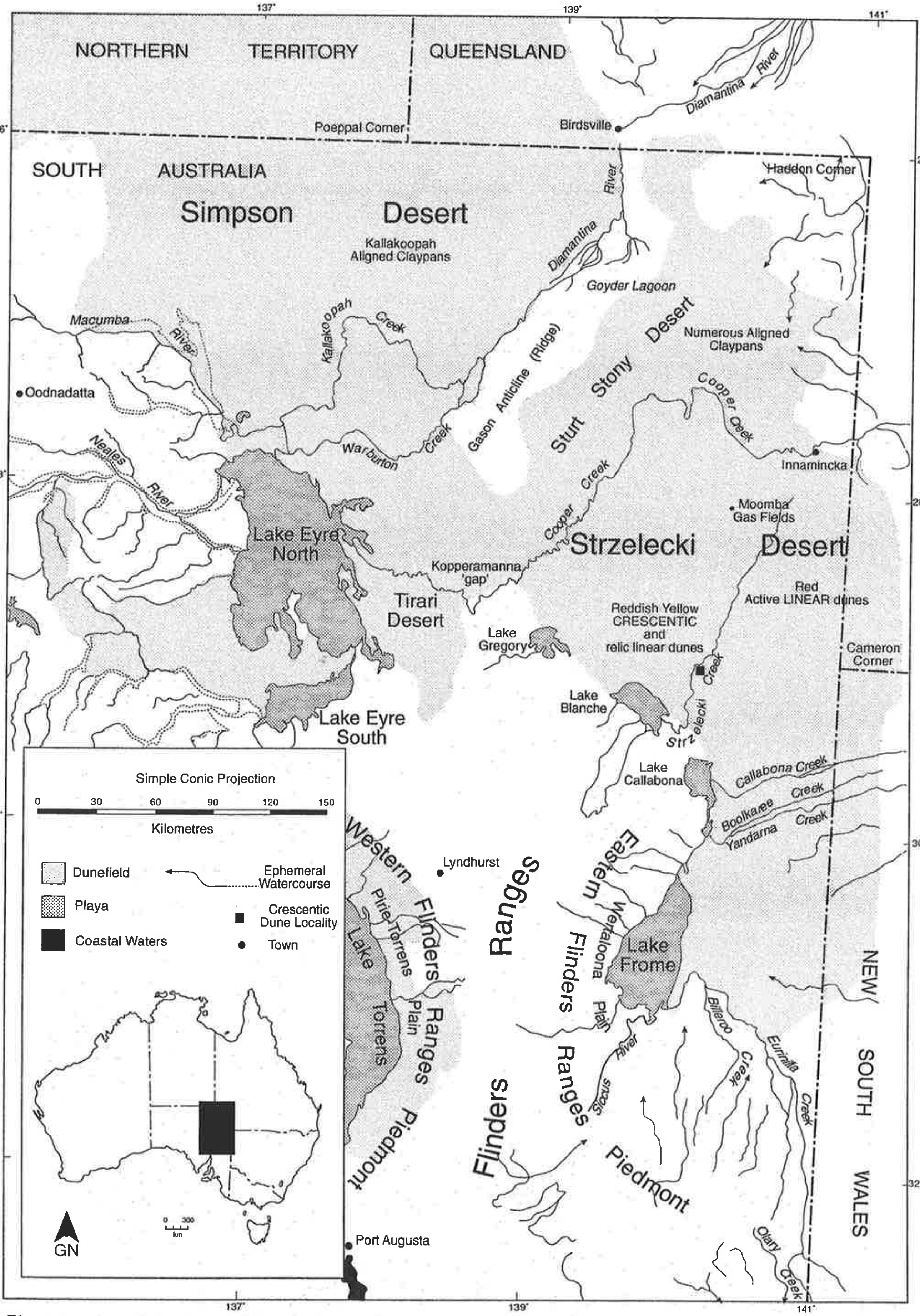


Figure 1.1 Regional setting of the Strzelecki - Simpson dunefields and major drainage patterns of NE South Australia.

Palaeocene, whilst epeirogenesis structurally deformed it into a series of anticlines and synclines in the early Oligocene to early Miocene. Contemporaneous down-warping produced two sedimentary basins in the Lake Eyre region of the Southern Simpson and the North Strzelecki dunefield and Lake Frome area (Wasson, 1983b). Within the Strzelecki basin, the Eyre Formation is overlain by the Namba Formation, a Miocene sequence of alternating, fossiliferous, fine to medium sands and thin dolomitic beds (Callen, 1977). The equivalent lithology of the Simpson basin, is the Etadunna Formation of lacustrine and fluvial limestone, dolomitic clay and interbedded silty sandstone. A period of epeirogenesis orchestrated downwarping in the Kallakoopah Pans and near surface expressions of the Miocene formations in the regions of Lake Eyre and Lake Frome, in the northern Simpson at the edge of the dunefield near Birdsville, and on the flanks of the Gason Ridge (Fig. 1.1). Uplift has also occurred around the margins of the basins and along the Gason anticline (Wasson, 1983b).

The post-Miocene Willawortina Formation in the southern region of Lake Frome (Fig. 1.1) represents a period of fluviatile deposition (Callen and Tedford, 1976). Continuous deposition of lacustrine clays within an ancestral Lake Frome has developed the Millyara Formation which still accumulates today. The ancestral clays inter-tongue with the Eurinilla Formation, an alluvial sequence younger than the Willawortina Formation. The shrinking of Lake Frome over time, has abandoned transverse shoreline ridges that are aeolian in their upper parts on the eastern side, with linear dunes extending eastward from these ridges (Wasson, 1983a).

In general, the northern Strzelecki Desert of the Moomba area comprises post-Etadunna-Namba Formation sediments that are significantly fluviatile with aeolian sands near the top of the stratigraphic column. The lithology of the Cooper Creek alluvium upstream of Innaminka and the Diamantina alluvium north of Birdsville, is a sandy mud overlying poorly-sorted fluvial sands (Veevers and Rundle, 1979; Wasson, 1983a, 1983b). Downstream of Innaminka, beds of alluvium (sands and muddy sands) were reported by Wasson (1983a), while Rust

(1981) interpreted the upper 50 metres of sediments near Moomba to have been braided channel sediments derived from upstream of Innaminka. The aeolian Coonarbine Formation rests upon the fluvial Eurinilla Formation. Four separate units are recognised within the Coonabine dunes (transverse and linear), each divided by carbonate segregation (Wasson, 1983a). 'Palaeosols' of carbonate nodules and rhizoliths indicate near surface levels when dune building had ceased or slowed.

Sediment sources for the dunes of the Strzelecki-Simpson dunefields are predominantly alluvial and lacustrine (Wasson, 1983a; 1983b). Dune sands are of two predominant Munsell colours, red-brown (5YR 5/8) and pale brown to pale orange (7.5YR 6/6). Boundaries between these colours are often gradational except along the northern Strzelecki creek where the boundary is sharply defined by the floodplain (Wasson, 1983a). Adjacent to the eastern shore of Lake Frome, a region of both red-brown dunes amongst pale brown dunes exists. The pale sands of lacustrine origin build transverse dunes that extend into linear dunes downwind, while the red-brown sands and dunes are most probably derived from the alluvium of the Eurinilla Formation. Calcareous palaeosols have given ^{14}C ages of *ca.* 15,000 B.P. - 25,000 B.P. (Wasson, 1983a; 1983b).

Pale-brown linear dunes derived from the pale-brown sediments of the peripheral transverse dunes occur on the north-eastern shore of Lake Frome (Wasson, 1983a). Both east of Lake Frome and east of Strzelecki Creek, red-brown dunes overlie the Cretaceous Winton and Tertiary Eyre Formations. Wasson (1983a) has recorded that approximately 75 kilometres SSW of Innaminka, red linear dunes have formed on top of a remnant of the Narconowie meandering palaeo-channel. Interdune outcrops and exposures of the marine Winton Formation and the overlying fluvial Eyre Formation in Strzelecki Creek, are separated from the dunes by colluvial sediments on the edges of the low hills in the north and alluvial sediments at the Narconnowie channel site. Elsewhere, a non-descript sandy clay of unknown origin is the most likely source of the dune sand for these red dunes (Wasson, 1983a). The pale dunes of the Strzelecki Desert lie within a structural depression

(Wofner *et al.,* 1974) that reflects Permian structures of the Cooper basin. The outcrops of Winton Formation both along Strzelecki Creek and to the east of it (but found some 200 m below the surface near Moomba) suggest structural control of the depression for deposition, and therefore, the existence of a flexure parallel to Strzelecki Creek (Wasson, 1983a; 1983b).

The northern part of Strzelecki Creek acts as a sharp boundary between the red-brown dunes and the pale-brown dunes (Wasson, 1983a). Dunes 100 km north and south of Cooper Creek form groups separated by floodflats (Wasson, 1983a). The floodflats are bounded on their northern, downwind side by large transverse dunes up to 60 m high, while linear dunes stem orthogonally from the lee slopes. Both morphologies contain several stratigraphic units of aeolian sand separated by weak carbonate pedogenesis (Wasson, 1983a). All stratigraphic units contain fine sand-size clay pellets. Bowler (1973) suggested that clay pellets are produced by salinization of floodflats during flooding, being later deflated by strong winds, a requirement needed to overcome hydroscopic attraction between pellets. Subsequently, these were re-deposited into source-bordering dunes. Crystallisation of salts within the floodplain sediments as water evaporates and the presence of a shallow but fluctuating water table, may have mechanically pelletised the sediments. Flocculation of a salinised, partially-montmorillinitic sediment would also aid pellet development (Wasson, 1983a).

The present level of the water-table in the Moomba area is approximately 40 m below the surface, so that both environmental and hydrological conditions have changed significantly since the main phase of dune building some 14,000 year B.P. (Wasson, 1983b). Wasson (1983b) has recorded that during much of the last 23,000 years, most dunes were adjacent to floodplains receiving mud-rich loads. This concurs with the pre-13,000 B.P. dune construction phases being coeval with the deposition of fluviatile mud-rich sands, and with the dunes being rich in clay pellets.

1.4 Climate and Meteorology of the Australian Dunefields

The major part of the dunefields lies within the 250 mm annual average isohyet. While most dunes are concentrated in regions that receive between 150 mm and 250 mm annual average rainfall, the centre of the arid zone north of Lake Eyre, Simpson dunefield, receives <125 mm annum^{-1} (Ash & Wasson, 1983).

Bowler *et al.,* (1976) propose a 14,000 - 25,000 year B.P. age (the last glacial maximum) for the linear dune building episodes of the central and northern deserts. This was a time of greater windiness and aridity, and lower global temperature (Oeschger, 1979). During this time, a dunefield whorl spreading through some 19° of latitude and some 30° of longitude, evolved over the heart of arid Australia. The anticlockwise nature of the dune ellipse has been associated with anticyclonic pressure cells (Sprigg, 1963, Mabbutt, 1967), although its extent may have been further north than it is today (Sprigg, 1963). Brookfield (1970) on the other hand, has shown that the current time of maximum anticyclonicity does not coincide with the time of high wind and greatest sand-moving potential, but better correlates with the southern movement of the northern cyclones. However, most authors have associated the formation of the dunes with the northerly shift of subtropical high pressure cells in summer.

In summer, two high pressure cells lie south of the continent. These slowly move northward and, in June - July, eventually centre themselves over Western Australia and the Eastern Great Dividing Range at about the 30° parallel. For this reason the dune pattern of continental Australia has been associated with the 'winter anticyclone'. Wind speed is greatest in col areas between the centres of maximum anticyclonicity, with the Simpson Desert lying to the north-west of the col zone. Nonetheless, Brookfield (1970) has ascertained that the closer the approach of the anticyclonic centre to the interior of the continent, the greater the decrease in sand-moving winds. Further data refuting the winter anticyclone mechanism of dune genesis, are the present day anticyclone dimensions which are approximately twice that of the dune ellipse. It is difficult to correlate these dimensions with outward wind flow and dune pattern. To produce sand-moving wind speeds required for dune development it would require much narrower anticyclones and steeper pressure gradients with very little seasonal latitudinal shift (Brookfield, 1970). Although this may have occurred in the Pleistocene, it does not occur with the current seasonal cyclonic cycle.

The incidence of cyclones in summer is greater than the anticyclones of winter (Brookfield, 1970). Two centres of maximum cyclonicity lie approximately over the 20° parallel, and are well correlated with summer and the highest frequency of sand-moving winds. Although the cyclonic mechanism does not totally explain the wind patterns that may have assisted in the formation of the dune pattern, it does display conditions required for maximum potential sand mobility. In general, cyclonic movement from the north of Australia towards lower central latitudes offers a better explanation of the current day wind strengths and patterns than does the movement of anticyclones from the south towards the north during winter.

Brookfield (1970) recognised the absence of a direct association of dune pattern with the present wind patterns, but also realised that these parameters do not significantly diverge from the current climatic belts. It is simplistic to model the dune pattern solely upon anticyclonic - cyclonic activity. The examination of dune trends and mean weighted wind direction (i.e. winds of sand-moving ability) have identified both significant negative (anticlockwise) and positive (clockwise) deviations[1] (Brookfield, 1970). Irrefutably, dune type as well as dune orientation is apparently influenced by the effect of relief and sand supply (Brookfield, 1970). Deflection or channelling of winds by higher relief, such as the Murchison-Davenport Ranges and the Harts-MacDonnell and Petermanns Ranges may be significant contributors to these effects, while in some instances, katabatic air flow may also be locally influential. The interaction of high and low pressure systems, topographical

[1] Negative deviation are those mean wind directions that lie more north or east than the mean dune direction, while positive deviations are those mean wind directions that lie more to the south or west of the dune trend (Brookefield, 1970).

divides and a series of dune forming events rather than a single time of origin, are mechanisms involved in forming the elliptical dune pattern of the Australian arid zone.

Fryberger (1979) using two meteorological sites (Oodnadatta and Giles) regarded the Australian dunefields as being environments of low to intermediate wind energy, that demonstrated little potential for sand transport. Using the Oodnadatta meteorological site (Fig. 1.2), Fryberger (1979) was able to demonstrate the sand moving capability (16 compass directions) for the monthly wind patterns of the southern Simpson Desert. Although this site is some 400 km west of the Strzelecki dune locality here studied, these results mirror the aeolian processes encountered in the Strzelecki Desert. Examination of the two dominant modes of the Fryberger sand roses[2] for each month, shows that during October through April the maximum sand drift potential is from southerly winds. During these predominantly summer and autumn seasons, saltation is towards the north, a prediction verified during the January-February field visit. During this time, significant saltation and dune advance was northerly, especially from mid-afternoon to the early morning of the next day. Conversely, the autumnal form of the crescentic dunes was seen to change very little and coincided with the meteorological summary of prevalent S and SSW winds. For May however, the meteorological data (Fryberger, 1979) shows variation of primary dominance to the SSW with secondary modes of S and SW, while June - September data demonstrate the existence of a prevalent northern vector coinciding with SW and SSW orientations. These results are in close agreement with both the winter and spring field observations of form reversal and easterly horn elongation. In addition, Brookfield (1970) has demonstrated that considerable variability of sand-drift potential exists in the central Australian deserts and that in many cases the sand-drift potential is site specific. For example, the highest sand-drift potential occurs at Finke on the NW side of the Simpson Desert, between low potentials at Oodnadatta and Alice Springs. Such regional

[2] A sand rose is a circular histogram which represents potential sand drift from 16 inter-cardinal - cardinal directions. Using WMO 'N-summaries' and a modified Lettau formula: $Q \propto V^2(V-12)t$, where Q is the annual rate of sand drift, V is the wind velocity (knots, at 10 m), t is the percentage of time during which the wind blew from a particular direction (expressed as percentage on the N summary), and 12 is the assumed threshold friction velocity for sand in the 0.25-0.30 mm diameter range, modes (vectors) of monthly *potential sand drift* can be illustrated.

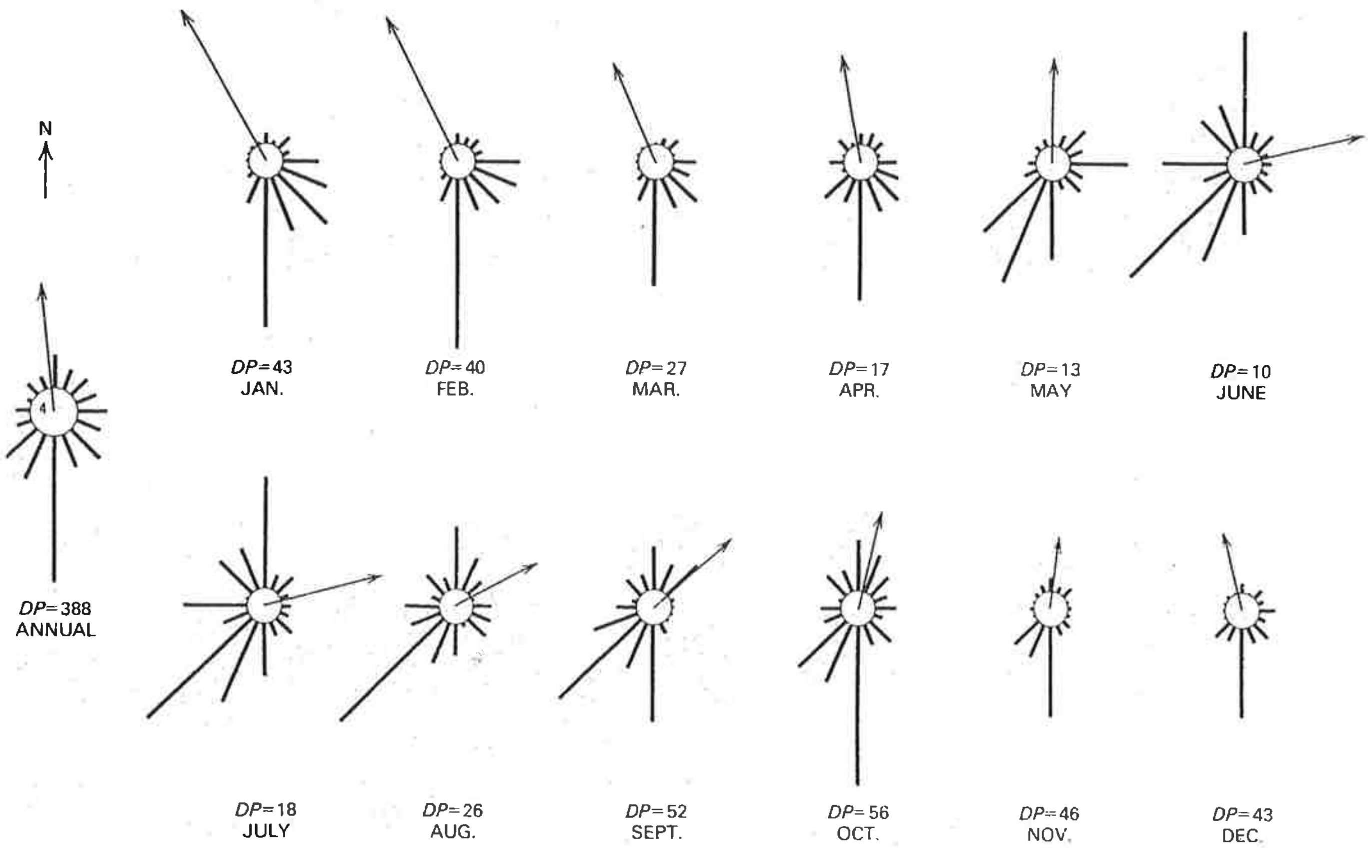


Figure 1.2 Annual and monthly wind roses for Oodnadatta, SW Simpson Desert, South Australia. A wide unimodal wind distribution of effective winds shown with the annual wind rose is a result of the interaction of winds between the SW and SE throughout the year. Arrows indicate the resultant drift direction. The number in the centre circle of the rose is the reduction factor or the number by which the vector unit total of each sand rose arm was divided so the longest arm would plot at < 50 mm. DP (drift potential, in vector units), is a measure of relative sand-moving capability of the wind. (from Fryberger, 1979)

variability is undeniably related, in part, to conditions at the locality (Brookfield, 1970; Ash & Wasson, 1983). These factors embody windiness, sand supply, topography, sand moisture and cohesion, and vegetation. Ash and Wasson (1983) have clearly demonstrated that the major part of the central Australian deserts lie in an area of low windiness, and it is this, rather than vegetative stabilisation of the dunes, that accounts for the low degree of sand mobility. For example, the present vegetation cover of the more arid parts of the Simpson-Strzelecki dunefields is not sufficient to stop sand movement. Significant sand mobility occurs with up to 35% plant cover (Ash and Wasson, 1983). To mobilise the Australian linear dunes, a minimum increase in windiness by some 20% - 30% is required (Ash and Wasson, 1983).

The diurnal character of the wind pattern for the southern Simpson Desert (Brookfield, 1970) is very similar to the qualitative observations made at the Strzelecki locality. Minimum wind speeds occur at 0600 hours with maximum speeds generally around midday, irrespective of the season (Brookfield, 1970). The afternoon speeds were relatively similar to the maximum midday mean, but by 1800 hours, wind speed had again significantly diminished. Mid-morning presented the most variable results, with low or high wind speeds being seasonally-dependent. Typically, the winter period produced velocities similar to the 0600 mean, with the summer 0900 readings often surpassing the 1200 hour mean maximum (Brookfield, 1970). In summary, it is clear that light winds to calms occur in the mornings of the winter months, with greatest wind speeds resulting from midday onwards. The highest velocities peak at the summer optimum.

As discussed in the preceding sections, current dune construction is not prominent in the modern Australian deserts, and is geographically restricted to areas of abundant sediment supply, high windiness and minimum vegetation. For this reason the genesis and current active nature of the Gurra Gurra (Strzelecki) crescentic dunes is somewhat enigmatic amongst the prevalent inactive, relict linear forms. Their presence reflects a relatively closed geomorphic system that is localised in extent, and very site specific regarding its

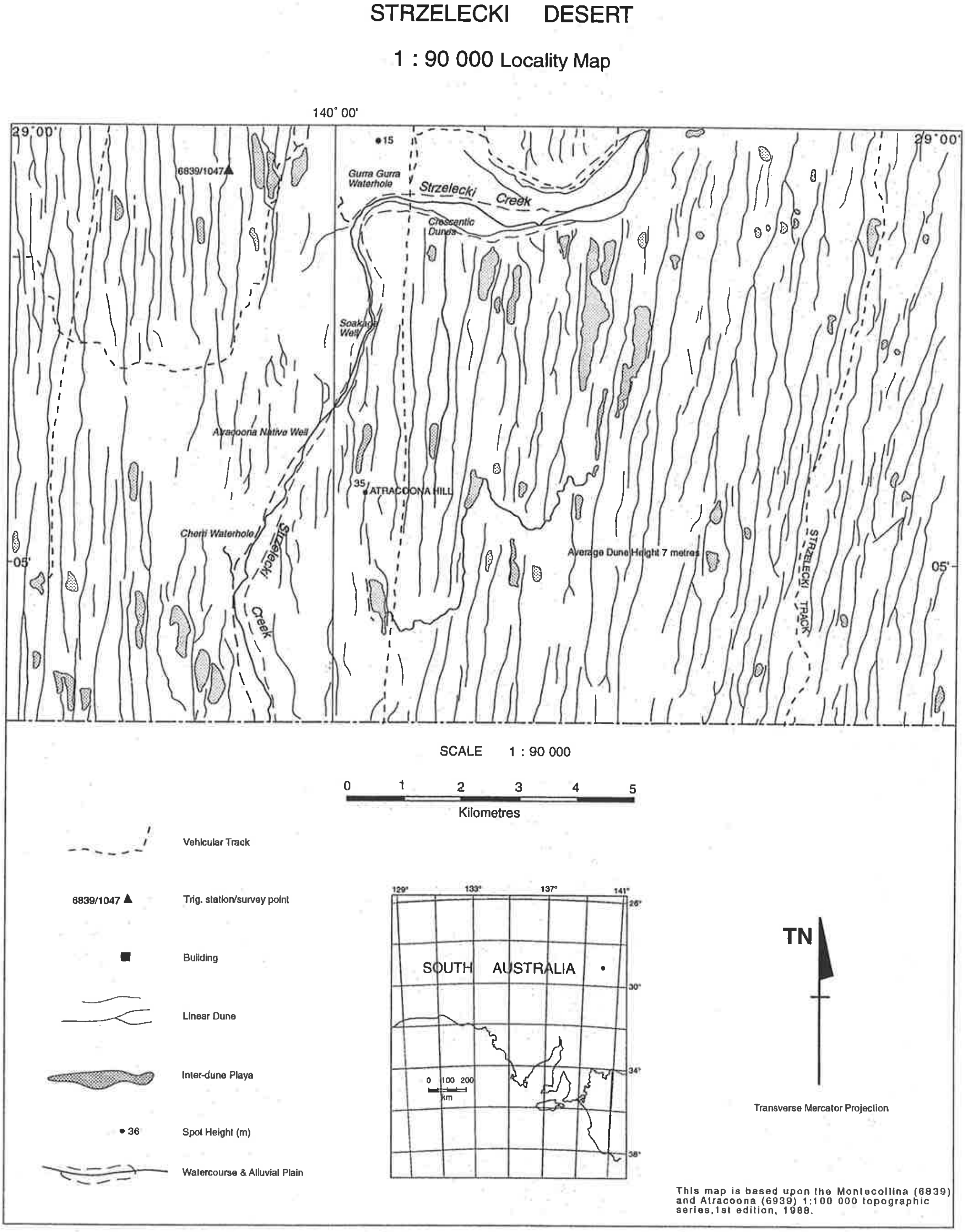
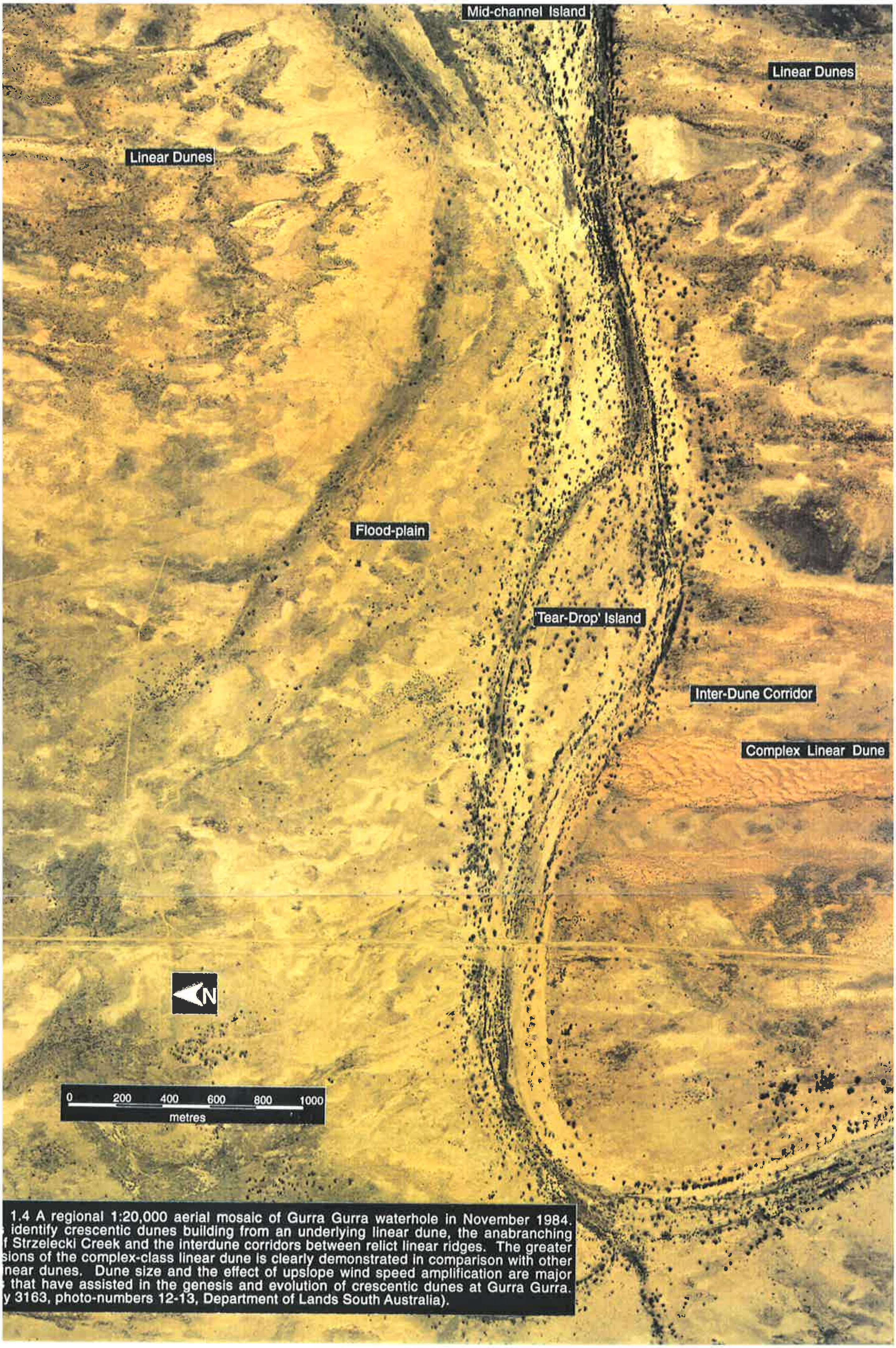


Figure 1.3 A 1:90,000 topographic-locality map of the crescentic dunes at Gurra Gurra waterhole, Strzelecki Desert, South Australia. Note the approximate north-south orientation and the prevalence of Y-junction openings towards the south for the inactive, simple linear dunes of the region.



1.4 A regional 1:20,000 aerial mosaic of Gurra Gurra waterhole in November 1984. identify crescentic dunes building from an underlying linear dune, the anabranching f Strzelecki Creek and the interdune corridors between relict linear ridges. The greater sions of the complex-class linear dune is clearly demonstrated in comparison with other inear dunes. Dune size and the effect of upslope wind speed amplification are major that have assisted in the genesis and evolution of crescentic dunes at Gurra Gurra. y 3163, photo-numbers 12-13, Department of Lands South Australia).

micro-meteorological and sand source conditions. It is the purpose of this research to explain the existence of the active Gurra Gurra crescentic dunes.

1.5 Research Objectives

The Strzelecki Desert's crescentic dunes of north-eastern South Australia (29° 01' S, 140° 02' E) dominate an approximate 2400 m x 320 m area, originating from the crestal zone of a quartz-sand linear ridge at a site known as Gurra Gurra waterhole (Fig. 1.3 and Fig. 1.4). The interaction of linear and crescentic crestal morphologies place the linear dune in the complex class-type. However, the identity of the linear morphology is nebulous due to parasitic erosion with contemporaneous re-deposition of the sediment into crescentic dunes, sub-parallel to the longitudinal axis of the underlying linear dune. These dunes show a quasi-progressive development of morphology from sand mound, barchan, barchanoid to transverse ridge and represent a local reactivation of only the linear dune upon which they are sited. All other linear dunes of the immediate area remain inactive, relict forms.

For this thesis, the research objectives explore the spatial and temporal changes and interactions of both gross dune morphology and dune surficial sedimentology for the crescentic dunes of Gurra Gurra waterhole in the Strzelecki Desert of South Australia. Specifically the focus is to examine the spatial variation of dune morphology using:

- dimensional differences of dune length, width, wavelength and height, compared to crescentic dunes in other terrestrial sand seas and the dunefields of Mars. Dimensional similarity proposes that dynamic similarity exists, and that the origin and evolution of the Gurra Gurra duneforms are allied to the process-response mechanisms of crescentic dunes found elsewhere.
- comparative micro-geomorphic change of element (lee, crest-brink, stoss) morphology in different seasons.

The exploration of surficial sedimentological change for dunes of similar size investigates the comparative sedimentology of:

- dissimilar micro-geomorphological elements and their surficial sedimentological distributions in each of the four seasons and,
- the surficial sedimentology of similar micro-geomorphic elements in all four seasons.

The examination of the spatial and temporal comparisons between dune morphology and sedimentology precedes:

- fundamental interpretation of genesis and evolution for the crescentic dunescape of Gurra Gurra waterhole.
- discussion of the likelihood of similar processes being active in other terrestrial and extra-terrestrial crescentic dune-fields.

1.6 Research Hypotheses

The research hypothesis is constructed so that it describes mathematical relationships for dune morphometry and dune sedimentology. These relationships are then used to understand the variation, processes and interactions between both morphology and sedimentology for the Gurra Gurra dunes over time.

The analysis of dune morphology requires both the null (H_0) and alternative - working (H_1) hypotheses to be constructed with the purpose of establishing the reliability of predicting the dependent variables from the regression equations for the morphometric parameters, of length, width, wavelength and height. The null hypothesis is that no linear relationships exist between two geometric variables, while the alternative hypothesis is that the dependent variable(s) *Y*, can be adequately explained by interaction with the independent variable(s) *X*. Regression analysis involves the use of correlation coefficients to establish the probability of the observed correlations, *(r)*, having arisen by chance. Formally, these hypotheses are given as:

- $H_0 : r = 0$ This reveals that there is no statistically significant relationship between the independent variables and the dependent variables. This is manifested in the total variance of Y being equal to the unexplained (residual) variance. That is, there is 'no explanation'.

- $H_1 : r \neq 0$ This is a statistically significant relationship between the independent variables and the dependent variables. This is manifested in the explained variance increasing at the expense of the unexplained variance. The degree of explained variance is significantly greater than that of the residual variance. The significance of 'difference' is given at the 0.05 level of significance whereby the observed r is significantly greater than or less than zero.

If rejection of the null hypothesis is identified, then the working hypothesis must be accepted.

Similarly, the statistical hypotheses used to determine the existence of spatial and temporal difference of dune surficial sedimentology, are established using both the Kruskal-Wallis and Scheffe type projection methods to test for 'difference' between the bivariate relationships of:

- dune morphology and sediment distribution
- sediment distribution and season.

This part of the research hypothesis contains two interrelated themes, involving spatial and temporal differences, that require individual alternative and null hypotheses. Firstly, it is necessary to establish whether there exists significant spatial variation of sediment descriptors for any of the distinct morphologic element(s), in respect of each season. Secondly, and irrespective of whether the first hypothesis is found to be true or untrue, it is necessary to determine whether there is a significantly quantifiable change in granulometric measures for specific morphologic elements that could be attributed to temporal change of wind regime.

Accordingly, the two-fold nature of the null hypothesis states that 'no difference' is perceived for dune sedimentology because of (i) the micro-morphologic location, irrespective of temporal setting (season), and (ii) that 'no difference' of the sediment distribution for similar micro-morphologic location is recognised, respective of temporal setting. Contrary to H_0, the alternative hypothesis accepts that (i) sedimentology is attributable to position on the dune, irrespective of season and (ii) that dune sedimentology is a consequence of seasonal influences irrespective of micro-morphologic location. Simply, the research hypotheses for the testing of sedimentological change can be summarised as:

- the null hypothesis where, H_0 : sedimentological difference = 0
- alternate hypothesis where, H_1 : sedimentological difference $\neq$ 0

It is perceived that the inter-relationships between morphology, granulometry and seasonal aeolian regime will assist in discerning a morpho-dynamic model for the Gurra Gurra crescentic dunes.

1.7 Dissertation Outline

The structure of this dissertation and the crux of the research objectives are found in:

- Chapter Two, which is a review of the global literature that is deemed most instructive to the structure of this research thesis. Referenced summaries of prior studies in aeolian geomorphological and sedimentological theory, show the fundamental knowledge base that is necessary for a study of this type.

- Chapter Three, which outlines the methodologies and the appropriateness of the techniques employed in the sampling, laboratory preparation and data analysis stages of the study. Because no single benchmark is established from which to study the interaction of form and sedimentology of desert dunes, reviews of the adequacies and inadequacies of individual methods are also supplied.

- Chapter Four, which examines the morphometric identity of the crescentic dunes using both aerial and field-derived measurements. All dimensional data are summarised using mathematical procedures of correlation and modelling by simple linear regression to evaluate the morphometric identity of the dunes compared with crescentic dunes in other terrestrial sand seas, specifically the barchans of the Pampa de la Joya of southern Peru, and the Salton sand sea of southern California.

- Chapter Five, which reveals how minor depositional (ripples, lee dunes) and erosional (scour flutes, yardangs) landforms of the Gurra Gurra crescentic dunefield assist in the determination of both regional and local wind directions. The accuracy of ripple orientation as indicators of wind direction is also discussed.

- Chapter Six, which demonstrates the sedimentary environment from which the dunes are derived. Both qualitative and quantitative discussion and comparative statistical analyses of granulometric distributions for some 300 sand samples are made between both different spatial and temporal parameters.

- Chapter Seven, which integrates the results of Chapters 4 and 5. The discussion of process-response mechanisms attempts to explain the variation of dune morphology. The interactions and interrelationships of the dunes with sedimentary features of erosional (yardangs, desiccation polygons) and depositional (lee projections, current shadows) origins elucidate a morpho-dynamic model from which the genesis and evolution of the Gurra Gurra crescentic dune-scape is made.

- Chapter Eight, which integrates the results of Chapters 4, 5 and 6. This chapter emphasises the process-response mechanisms responsible for surficial sedimentological variation of the dunes in relation to dune morphology and aeolian regime.

- Chapter Nine is the conclusion and presents an overview of the morpho-dynamic model of genesis and evolution for the dunes of Gurra Gurra waterhole, with subsequent discussion concerning the actions of similar processes in affecting the form and sedimentology of crescentic dunes in other arid settings. A review of future research directions is also presented.

It is hoped that the discussion of the Strzelecki - Gurra Gurra dunescape will augment the knowledge base from which a more complete understanding of aeolian bedforms may be derived. The distribution of dunes and ripples, yardangs, lee projections, nebkha and desiccation polygons belongs not to one arid landscape either terrestrial or extra-terrestrial, but in association with numerous desert regions. It is hoped that the observations and discussions given here, benefit the understanding of the associations between gross morphology and sedimentology for an environment in which spatial and temporal variation is unrelenting.

2 Terrestrial and Extra-Terrestrial Dune Studies

2.1 A Historical Perspective of Desert Dune Studies

The quantitative study of desert landforms is a relatively recent science. Many of the early studies were of a descriptive nature geographically documenting the landforms of a particular locality. For example, the investigators Von Tschudi (1946) and Raimondi (1929) examined the orientation and overall morphology of the barchan dunes of southern Peru (Finkel, 1959; Hastenrath, 1967), but with a notable absence of discussion concerning the origin and evolution for these dune forms. Conversely, Madigan (1936) provided observational insights into the origin and evolution of Australian sand ridges (linear dunes), but lacked supportive physical principles for his model. Prior to the pioneering work of Bagnold (1941), comparative studies of dune development for different localities was negligible, while actual analysis of sedimentary materials and aeolian processes was virtually absent.

Bagnold (1941) attempted to unify all the elements of desert studies into a cohesive, quantitatively deduced set of principles, reflecting not just his studies in the Sahara, but also the observations of others around the globe. This work has become the bench-mark for much of today's understanding of arid zone geomorphology. Through the interdisciplinary affiliation of geology, meteorology and physics, evolved an understanding of the mechanisms involved in the origin and evolution of the terrestrial sand seas and associated desert landforms.

Another significant development in the study of global desert terrains was brought about by satellite imaging and greater vehicular access. From this alone, the study of aeolian geomorphology has proliferated in the last few decades. From global comparisons of the Earth's arid zones, McKee (1979b) has developed a scheme of classification for the various dune morphologies, accompanied by an account of the processes involved in sculpturing aeolian landforms. Subsequently, this planetary approach has aided in correlating global wind regimes with dune morphologies, ensuing an understanding of the terrestrial distribution of aeolian landforms.

These same techniques of analysis have also been applied to the study of extra-terrestrial deserts. The missions of Mariner 4, 6, 7 and 9 (1965-1971) and Viking 1 and 2 (1976) have revealed vast regions of aeolian landforms within the Hellespontus and north polar erg localities of Mars. Voyager in the 1989 encounter with Neptune's largest satellite Triton, imaged wind streaks emanating from craters of ambiguous origin (Croft, 1990), while the radar mapping of Venus by the spacecraft Magellan has revealed obstacle-induced wind streaks and possible dune forms (Weitz *et al.*, 1991).

The subsequent sections of this chapter review specific topics of interest that are required in the research of the spatial and temporal geomorphology and surficial sedimentology of duneforms at Gurra Gurra waterhole. Topics of review have therefore, purposefully concentrated on dune morphology and morphometry, dune surficial sedimentology, comparisons of dune types on Earth and Mars, as well as, the processes involved in dune genesis, evolution and development of sedimentological character.

2.2 Aeolian Landforms: Morphologies and Origins

Dunes are constructional topographic forms that develop from assemblages of loose fragments of mineral or rock in specific size fractions under the action of a moving fluid, such as the wind. Bagnold (1941) gave the first quantitative insights into this phenomenon, recognising that dune-forming sands predominantly move by saltation. Sand grains travel in a downwind ballistic trajectory, on assuming the motion of the wind, and impact the surface at a shallow angle. If this impact occurs on much larger particles or upon a solid surface, the grain rebounds, although it may impart some motion (traction) to the larger particles. If it encounters particles of a similar size or smaller, and if the wind is strong enough, it will spray them into the air. At low wind velocities the motion can be sustained only on rock and pebble surfaces because the energy of impact into a surface of fine

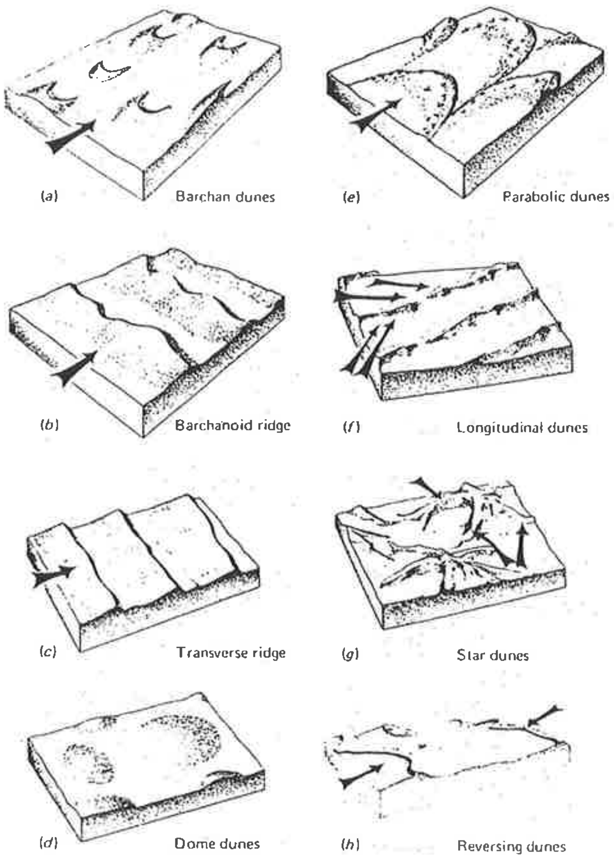


Figure 2.1 Illustrations of the principal dune types and the wind vectors responsible for their formation (from McKee, 1979)

Dune Type	Morphology (Plan)	Slipfaces	Wind Regime
Barchan	crescentic	1	unimodal, transverse
Barchanoid	coalesced crescentic	1	unimodal, transverse
Transverse	asymmetrical ridge	1	unimodal, transverse
Linear	symmetric ridge	2	wide unimodal, parallel
Reversing	asymmetric sinusoidal	2	bimodal, transverse
Parabolic **	U shaped	≥ 1	unimodal, parallel
Star	central peak, 3 arms	≥ 3	multiple
Dome	circular or elliptical	0	?
Sheet *	sheet	0	?
Streak *	linear streak	0	unimodal, parallel

Table 2.1 Characteristics of the common dune morphologies

* Not all morphologies are true dunes displaying a 3D aspect, nonetheless, they are products of aeolian deposition.

** A form which attributes its morphology to the stabilising influence of vegetation and/or moisture, more than wind strength and direction. Parabolic dunes are fundamentally a deflation feature where deposition is limited to the migratory margins (McKee, 1979).

particles becomes dissipated. Thus under these conditions, moving grains collect on patches of stable grains, and the landscape is not uniformly mantled but is comprised of segregated bodies of sand (Bagnold, 1941).

The classification of aeolian landforms is primarily based upon two descriptive criteria, (McKee, 1976) namely the shape and the number of slip faces or lee side surfaces. Both these characteristics can be determined from ground, aerial and orbital platforms of observation. McKee (1976) devised a classification of dune morphologies into three distinct types using a global comparison of terrestrial deserts. These morphological groups are:

- Simple - an individual structure showing a form that characterises a dominant wind direction and strength that is portrayed in the number of slip faces.
- Compound - where two or more dunes of the same morphology coalesce or superimpose upon each other.
- Complex - where two or more different dune morphologies coalesce or develop upon each other or within close proximity to each other, such as within interdune corridors.

Transitional forms of these morphologies are numerous and are representative of :

- wind strength, direction and duration
- sediment supply
- obstacles encountered by the saltation paths of the sediment
- moisture content of the dune bodies
- vegetative stabilisation of the sand mass.

Combinations of two or more of these factors direct the evolution of the morphological class (McKee, 1976). The dune morphologies that are most commonly observed in the terrestrial sand ergs are as shown in Table 2.1 and Fig. 2.1.

2.3 Physics of Dune Processes

2.3.1 Wind-Bed Interactions

The three principal modes of aeolian transport are, suspension of particles < 60 - 70 mm, saltation of particles between 60 - 1000 mm and creep via saltation impact for sediment > 500 mm in diameter. Both wind tunnel and field studies have demonstrated that wind flow of a given threshold velocity over a loose, dry surface free of obstructions, will dislodge and entrain sedimentary particles of sand, clay and silt. Wind velocities near the surface are high and develop shear forces across the surface, and are expressed in terms of drag or frictional velocity or *shear velocity* (u_*) :

$$u_* = \sqrt{\tau_0/\rho_a} \tag{2.1}$$

where :

t_0 is the shear force per unit area on the bed (gm $cm^{-1}sec^{-2}$)

r_a is the density of the air (gcm^{-3})

u_* has the dimension of velocity (ms^{-1})

In a more precise sense, however, grain motion occurs at the fluid threshold shear velocity, i.e. "...when fluid force (lift, drag, moment) exceeds the effects of the weight of the particle and cohesion between adjacent particles" (Lancaster and Nickling, 1994 p. 450). The fluid threshold shear velocity is expressed as:

$$u_{*_t} = A\{[\rho_p - \rho/\rho]gd\}^{1/2} \tag{2.2}$$

where, A is an empirical coefficient dependent on grain characteristics and equal to ~0.1 for Re_p > 3.5, r_p is the grain density, g the acceleration of gravity and d the grain diameter (Bagnold, 1941).

Hence, saltation leads to the transfer of momentum from the saltating grain to the stationary grains of the surface. This results in the impact energies ejecting surface grains into the air stream, but at a lower shear velocity than that by fluid forces. A new lower threshold of motion, the impact threshold, is acquired (Lancaster and Nickling, 1994). The grain-size lofted (fluid threshold) under a given shear velocity has been experimentally determined for different planetary atmospheres of different compositions, density and gravity (Iversen *et al.,* (1976b); Iversen and White, (1982); Greeley and Iversen, (1987); Leach *et al.,* (1989)). These results are illustrated in Fig. 2.2. When u_* induces saltation, the wind loses some of its momentum, changing the wind profile to non-logarithmic at this point *(z')* , but increasing the shear velocity (Bagnold 1941; Lancaster and Nickling, 1994). However, at a given height *(z')* above the surface (2 - 4 mm), wind speeds are unaffected and behave log-linear with increasing height. The parameter *(z')* may represent the mean saltation height of uniformly sized grains, with the saltation layer being estimated to be (z'.10) (Lancaster, 1994), but dependent on the 'hardness' of the pavement (Sharp, 1964). In a neutral, stable, stratified atmosphere with fully turbulent flow (high Reynolds number, $Re_p \geq 70$), velocity will increase logarithmically with height above the surface. Shear velocity is therefore related to the slope of the velocity-log-height curve (Cooke *et al.,* 1993), and is characterised by the Prandtl-von Karman equation:

$$\frac{u_z}{u_*} = \frac{\left(\frac{1}{K \ln z}\right)}{z_0} \qquad 2.3$$

where:

u_z is the wind velocity at height z

u_*. is the shear velocity

K is a Karmans constant (0.4)

z_0 is aerodynamic roughness length of the surface

2.3.2 Effects of Surface Roughness on Entrainment

The most fundamental process of dune origin is determined by a change of a terrain's surface roughness (Cooke *et al.,* 1993). Wind flow over a flat surface will result in a shear velocity decrease when the wind flow crosses a contrast of surface roughness, thereby

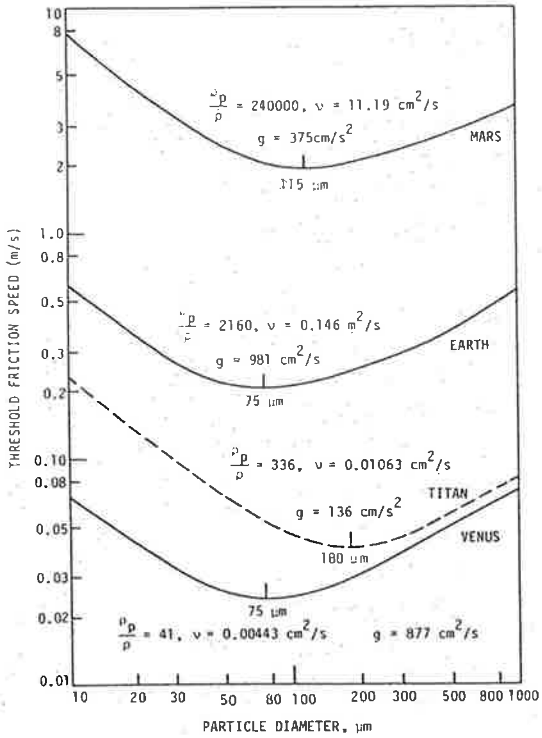


Figure 2.2 Comparative threshold friction speed versus particle diameter predictions for Venus, Titan, Earth and Mars. (from Greeley & Iversen, 1987)

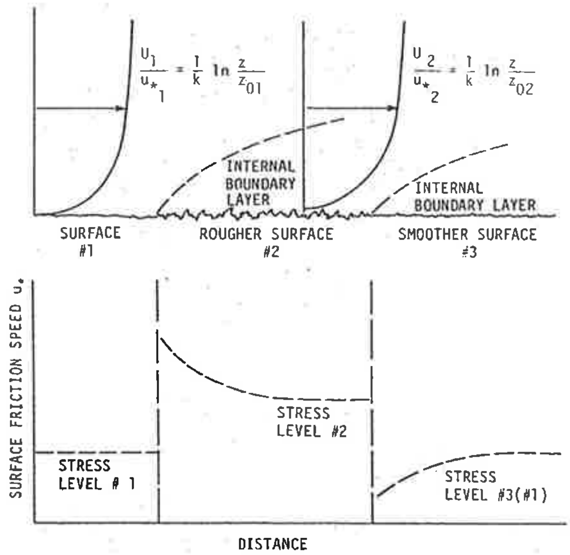


Figure 2.3 Schematic diagram of wind speed profiles (upper figure) and surface shear stress distribution (lower figure) associated with discontinuous surface roughness. (from Greeley & Iversen, 1987)

leading to the deposition of sand in aeolian transport. Also, a decrease in shear velocity will occur by air flow over small hollows which will similarly lead to sand deposition and accumulation at that point (Kocurek, 1990). When a dune is initiated in this manner, an apparent equilibrium size is achieved, from which the juvenile dune can advance from the template. This then allows other dunes to develop and migrate from this point of genesis (Cooke *et al.*, 1993).

If the surface is rough, u_* is higher in the same ambient wind speed relative to a smoother surface (Bagnold, 1941; Greeley and Iverson, 1985; 1986). Greeley and Iverson (1986) found that for discontinuous surface roughness in the Amboy lava field of the Mojave Desert, shear stress variation is found between the coarser basaltic and smoother alluvial surfaces. Because the rougher surface will eventually cause a decrease in the wind speed gradient near the surface, the temporarily high gradient from an upwind smoother surface is the cause of a temporary increase in surface shear stress (Fig. 2.3). The ability to erode is therefore greatest at the leading edge of the rough surface rather than further downwind. Where the converse exists, rough to smooth, the stress level at the leading edge is less than further downwind and a net deposition is expected (Greeley and Iversen, 1985; 1986).

Surface obstacles and variation in surface thermal conditions are also variables that affect shear velocity. Roughness elements such as vegetation, or different lithological surfaces, e.g. *aa* basalts, pahoehoe basalts, playa or gibber-plain have different aerodynamic roughness (z_0) values. Therefore a high density of roughness elements displaces the wind velocity profile upward to a zero-plane adjustment height (d_0) (Lancaster and Nickling, 1994). The wind velocity profile equation then becomes:

$$u/u_* = 1/K ln[(z - d_0)/z_0] \qquad 2.4$$

It is apparent that the multi-related processes of threshold motion and sand transport are complex and often difficult to measure and understand at the granular or micro level. The following section deals with the dynamics at the meso-scale and explores some of the principles that control dune morphology and surficial sedimentology.

2.3.3 Dune Dynamics

"Dune initiation, growth and equilibrium morphology of all dunes are determined by their sediment budget, defined as the balance between local erosion and deposition, summed for the dune as a whole" (Lancaster, 1994 p. 483). In obtaining equilibrium of form and migration, a meso-dune conserves the mass of sediment totalling its form, as well as the conservation of its morphology. This has been adequately demonstrated by both Finkel (1959) and Hastenrath (1967; 1987) for the barchans of Pampa de la Joya, Peru, and by Long and Sharp (1964), for the Tule Wash barchan of the Salton sand sea. Such attributes are described by the partial derivatives or sediment continuity equation (Middleton and Southard, 1978), as follows:

$$\frac{\partial h}{\partial t} = \frac{-\partial q}{\partial x} \qquad 2.5$$

where, h = local bed elevation, q = local volumetric sediment transport rate in the direction (x) and time (t). In other words, dune morphology is controlled by changes in sand transport rates. That is, wind velocity and hence u_*, determine the spatial patterns of erosion and deposition (Lancaster, 1989). Of fundamental importance in the understanding of dune morphology is the premise that the shape of the dune act as an obstacle to airflow in the planetary boundary layer. Such an obstruction leads to the compression of airflow lines at the crest of a dune, with an inherent increase in u_* upon the windward slope (Lancaster, 1989; Burkinshaw *et al.*, 1993), and is similar to the meteorological models of air flow over hills (Jackson and Hunt, 1975; Pearse *et al.*, 1981). Jackson and Hunt (1975) show that the magnitude of wind speed-up or amplification is expressed as:

$$\Delta s = 2h/l \qquad 2.6$$

where h is the height of the obstacle (dune, hill etc.) and l is the length measured parallel to the wind at $h/2$. A strong correlation between dune height and Δs shows a positive proportional increase between these two parameters. The amplification of wind velocities on the windward flanks of dunes increases the potential sand drift over that dune towards the crest. The fractional speed-up ratio as modelled shows that for *steep* dunes the average wind speed relative to upwind areas will increase up the flank to 280% at the crest and decline to 20 - 40% on the lee slopes (Jackson and Hunt, 1975; Lancaster, 1989). The magnitude of the increase however varies with the velocity, direction of the incident wind on the flank and height of the dune. (Lancaster, 1989). Amplification being greatest when the wind is normal to crest. Hence even at low wind velocities on the interdunes and lower flanks, where negligible or no sand transport occurs, upslope amplification of wind velocity can initiate vibrant crestal activity (Lancaster, 1989) As wind velocities increase on the lower dune-interdune, the ratio of wind speeds between crest and flank are less, and the sand transport rates decline exponentially. As demonstrated by Lancaster (1989) for linear dunes of the Namib Desert, the ratio between plinth and crestal transport potential decrease from around 200 at just above threshold velocity, to between 4-15 on a plinth with wind velocities of 11 m sec^{-1}. Hence as overall wind velocities increase, the magnitude of potential sand drift increases and transport becomes more uniform over the entire dune, with smaller relative crestal velocities and rates of transport.

Therefore increases in both wind velocity and shear velocity *(u_*)* upon the stoss derives increased transport rates towards the crest. Maximum transport is obtained at wind velocities just above the fluid threshold shear velocity, but decreasing as wind velocity exceeds u_{*t}(Lancaster, 1985a). Leeward of the crestline air flow expands and can also undergo separation (Lancaster and Nickling, 1989). Air flow separation can be of three styles:

- separated
- attached
- attached-deflected (Sweet and Kocurek, 1990).

The type of leeward flow is controlled by dune shape (aspect ratio), incidence angle between primary wind and crestline, and the stability of the atmospheric boundary layer. High strength, attached leeward winds generally associate with low aspect ratio dunes and oblique principal winds, while flow separation and weak, reversing eddy currents associate with dunes of high aspect ratio (Kocurek *et al.*, 1992). According to Lancaster (1989; 1985a), wind velocities over crescentic dunes recover downwind to the equivalent upwind velocities, at a distance some 10 times the dune height, and may assist in determining the spacing of crescentic dunes. To further illustrate the complexities involved in the form-flow interaction, Lancaster (1994) commented that seasonal multi-directional wind regimes and the presence of significant lee-side secondary flows are also of utmost importance in defining dune morphology and dynamics.

2.3.4 Threshold Velocity and the Effect of Slope

Transport processes (threshold velocities for sand movement and transport rate) on dune slopes in the NW Algerian Sahara have been examined by Hardisty and Whitehouse (1988). Using a Bagnold-style expression, bedload transport rate on horizontal beds can be derived using;

$$j_b = k(u^2 - u_{cr}^2)u \qquad 2.7$$

where, j_b is the mass transport rate per unit width, u is the wind speed at some characteristic height above the bed, u_{cr} is the critical or threshold velocity for sediment movement and k is the coefficient of proportionality related to the properties of the fluid and the sediment. However, on dune slopes the expression is corrected to;

$$j_b = Ak(u^2 - Bu_{cr}^2)u \qquad 2.8$$

when, $A = k_b/k_0$ and $B = uc_{rb}\text{-}u_{cr0}$ which are the ratios of the inclined-bed to flat-bed values (equal to unity on a horizontal surface).

Theoretical analysis of the effect of bed-slope on the threshold *B*, has been given by Allen (1982):

$$B = \frac{u_{crb}}{u_{cr0}} = 1.373 \sin^{1/2}(i-b) \qquad 2.9$$

where, *i* is the angle of internal friction of the sediment, (i.e. the angle of repose 32°) and *b* = the bed slope. The empirical work of Hardisty and Whitehouse (1988) gives an extremely good correlation with theoretically determined values. Theory predicts that u_{crb} (threshold velocity of a slope) would tend towards zero on a negative slope equal to *i*. Hence equation (2.7) is a good prediction of slope-inclusive thresholds, and shows that greater threshold velocities are required for sediment movement on positive (windward) slopes relative to lower velocities of negative (lee) slopes. Similar results relate the effect of slope transport on the threshold of motion and its application to the orientation of winds ripples (Howard, 1977).

2.4 Crescentic Dunes

2.4.1 Morphology and Morphometry

Crescentic dunes have been the subject of much research (Bagnold, 1969; Lettau and Lettau, 1969; Hastenrath, 1967, 1987; Finkel, 1959, 1961; Tsoar, 1974; McKee, 1966; Norris and Norris, 1961; Norris, 1965, Breed, 1977; Cutts and Smith, 1973; Breed *et al.*, 1979) for a variety of both terrestrial and extra-terrestrial terrains. Crescentic dunes include barchans and ridges of arcuate segments, termed barchanoid, often formed by the merging or coalescence of barchans. Barchan dunes develop where sand throughflow rates are high or where sand is limited in a predominantly unidirectional wind regime (Thomas, 1989). Where a secondary oblique wind direction is present, barchans will also develop, but with a preference for elongation of one of the horns in the general direction of the vector sum of the two winds (Tsoar, 1984). Barchans have a maximum height that is one-tenth their width (Mabbutt, 1977). Morphology changes with size, with smaller forms being flatter with reduced angles between the stoss side and desert floor (Hastenrath, 1987). It appears

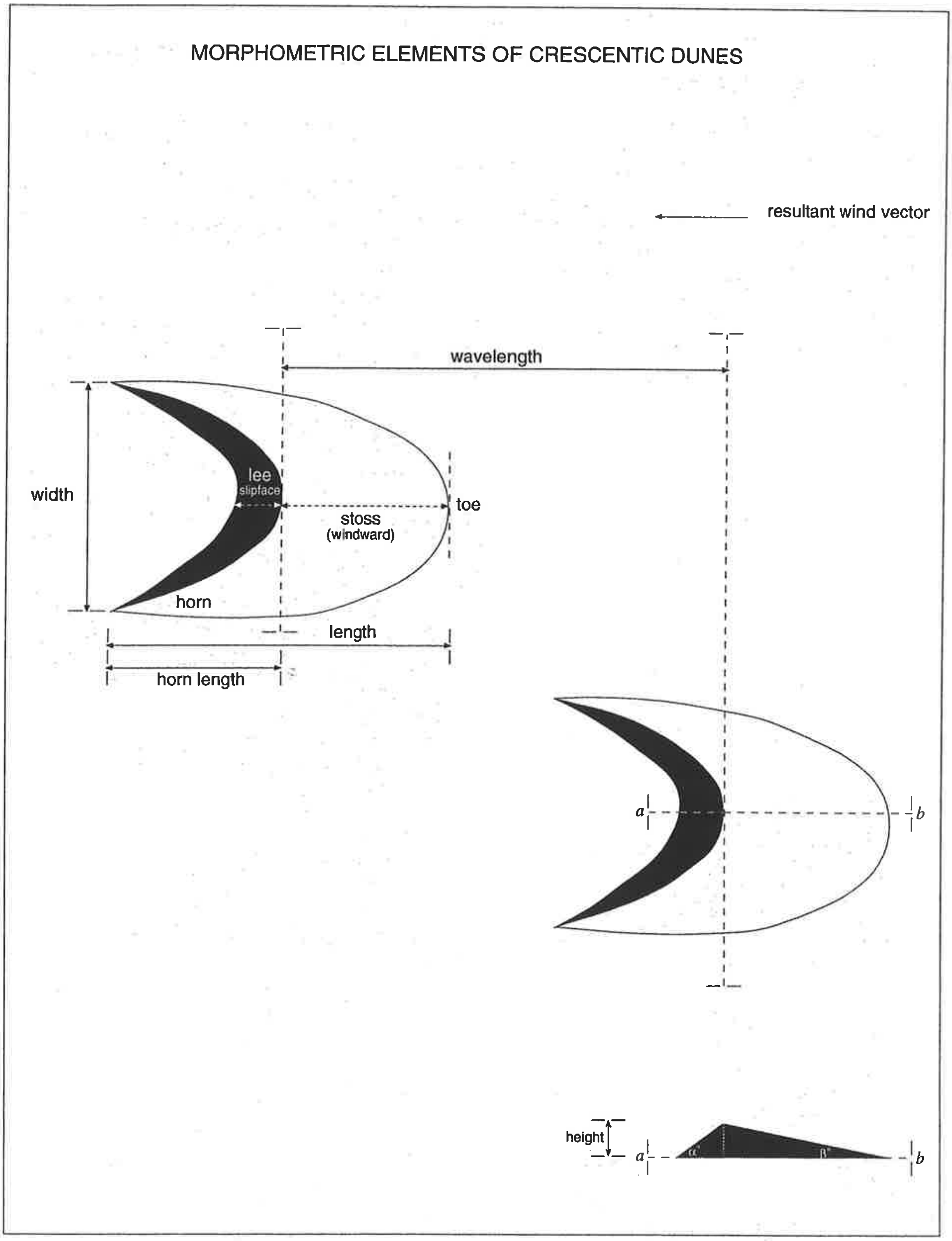


Figure 2.4 Morphometric nomenclature of crescentic dunes (not to scale). Angles of the lee and stoss slopes are designated alpha° and beta° respectively.

that size, and hence morphology, may be a function of factors such as sand supply, wind strength, atmospheric motion and age of the dune (Howard *et al.*, 1978).

With increasing sand supply, individual barchan dunes link or coalesce to form compound barchanoid ridges or more amorphous transverse ridges. This succession of barchan, barchanoid and transverse morphology is a genetic path controlled predominantly by sand supply, with the ridge-like morphology being more common than the individual barchan (Breed and Grow, 1979). A fourth type of crescentic dune is the mega-barchan; a term applied to compound forms of superimposed barchan dunes of different sizes (Simons, 1956; Norris, 1966). According to McKee *et al.*, (1977) and Breed (1979a) these dune-forms account for approximately half the dunes on Earth. Simple morphologies display widths of a few metres to half a kilometre and are found in most terrestrial deserts, while the larger compound varieties occur mainly in closed basins such as the central Sahara, South Eastern Arabian Peninsula and central Asia (Breed *et al.*, 1979). Crescentic dunes can vary in size by three orders of magnitude in the terrestrial environment (Breed *et al.*, 1979a; McKee *et al.*, 1977). However, scale ratios derived from the planimetric measurements of width, length and wavelength (Fig. 2.4) are very similar for both simple and compound dune morphologies.

Parameters proven useful in the determination of dune geometry are: length, width, wavelength, height, horn length, slope of the lee side or slipface and slope of the stoss side or windward face (Fig. 2.4). Finkel (1959) and Hastenrath (1967; 1987) made a very complete morphometric survey of the Pampa de La Joya barchan dune field of southern Peru. Similarly, Breed (1977) performed a global analysis of many sand seas with crescentic dunes for the purpose of comparison with dunes of the Hellespontus region on Mars. Using dune morphometry, shape can be compared throughout the terrestrial deserts by the method of dimensional analysis (Strahler, 1958). Breed's studies of both terrestrial and extra-terrestrial dune-forms using the measurement of geometric parameters, have proven useful in defining mathematical relationships between morphometry and morphology. The combined data sets of 21[1] terrestrial ergs and two Martian localities (Breed, 1978; Breed and Grow, 1979), demonstrate exceptionally strong coefficients of correlation at the 0.01 level of significance (Fig. 2.5). These infer strong and possibly causal relationships between morphometric parameters, although the designation of which variable is dependent on which, is not discernible from the dimensional elements used here.

For the means of width versus length, a Pearson correlation coefficient of r = 0.90 and the regression equation,

$$W = 0.15 + 1.50L \tag{2.10}$$

demonstrates a strongly significant ratio of length:width = 1:1.5 at the two tailed 0.01 level of significance.

Similarly the relationship between the means of wavelength versus length is also significant at the 0.01 level with a correlation coefficient of r = 0.91 and the regression equation:

$$WL = 0.24 + 1.17L \tag{2.11}$$

This implies that the spacing of crescentic dunes does not differ from zero as a result of chance, but rather that dune wavelength varies directly with dune length, at a ratio of length:width = 1:1.2.

Likewise the correlation of wavelength with width, shows a significantly strong correlation coefficient at the 0.01 level of significance of r = 0.80, while the equation,

$$WL = 0.37 + 0.62W \tag{2.12}$$

demonstrates a ratio of width : wavelength = 1 : 0.6.

These data show that the aeolian processes by which dunes develop are relatively similar (although the variables of these processes may be of different magnitudes) across local, regional, global and planetary scales.

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[1] Note that the following regression equations and scatterplots are the result of combining in this study, the data of Breed (1978) and Breed & Grow (1979).

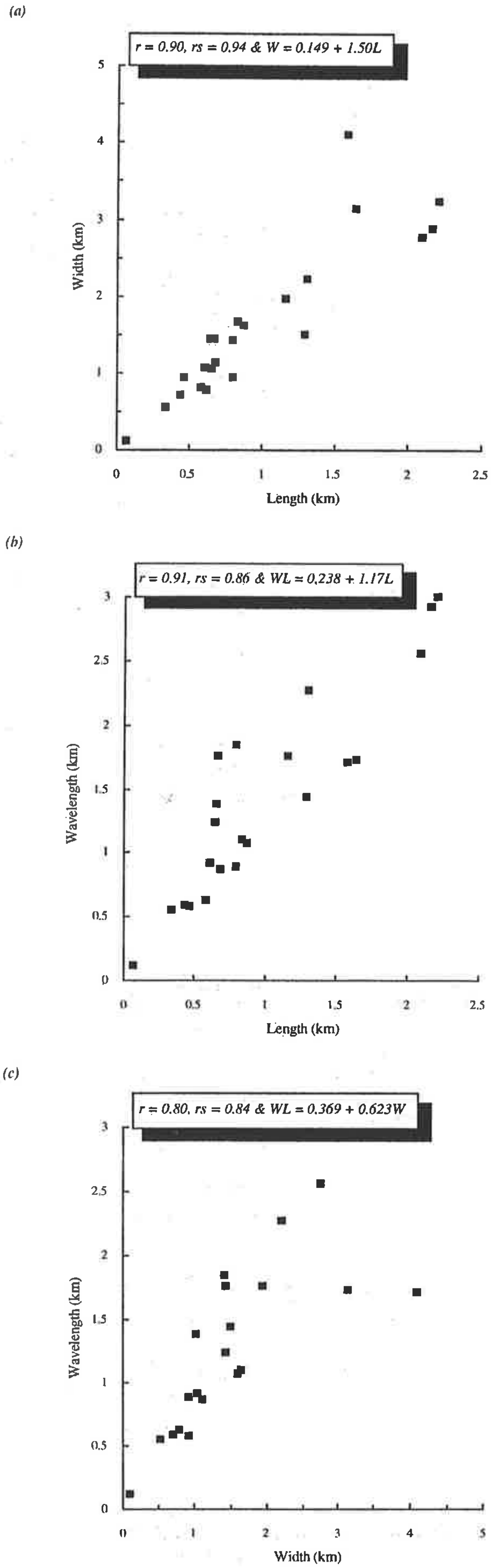

Figure 2.5 Bivariate scattergrams of length *(L)*, width *(W)* and wavelength *(WL)* for *(n = 23)* sand seas of terrestrial and extra-terrestial origin (data from Breed, 1977 and Breed and Grow, 1979).

Similarly, Long and Sharp (1964) classified the Salton dunes in California, U.S.A. according to their shapes, employing the ratio of horizontal distance along the axis of symmetry, from the trailing edge of the dune to the upper edge of the slipface *(a)*, and the distance between the tips of the horns *(c)*. Such nomenclature is equivalent to a ratio of length-to-width as shown in Fig. 2.4. The Salton dune group displayed considerable morphometric variety in which the length to width ratio ranged from >1.0 to < 0.25. No clear relationship between this ratio and other characteristics of the dunes was reported. A similar study of an individual member of the Salton Sea dunes (The Tule Wash barchan) was conducted by Norris (1966), where volumetric and advance rate studies were calculated. It was found that the Tule Wash barchan remained a relatively slim form, *a/c* ratio ≤ 0.25, (Long and Sharp, 1964), despite volume fluctuations of some 40%. This implies that this ratio is governed by factors other than sand flux (Norris, 1966), and may include subtle topographic influences or undetermined variations in the width or strength of the oncoming sand stream.

2.4.2 Origin

The saltation paths of sand grains over a solid pavement are reduced and slowed if they encounter patches of softer surface, such as accumulated sand (Bagnold, 1941). If the rate of sand deposition at these points exceeds the rate of sand erosion, sand accumulation and dune development occur at this soft surface point. Deposition is greatest on the lee side of a dune where wind velocities are lower and the flow pattern divergent, thus enabling a slipface to develop as the angle of repose is reached (Thomas, 1989). Sand accumulation is less at the edges at the site of deposition, as saltation rates are greatest here, leading to the development of barchan horns. The slipface moves forward more rapidly at the edges where the dune is lowest, creating a concave profile (Howard *et al.*, 1978).

Migration rates are related to the size of the dune, where small dunes advance more rapidly than larger forms (Watson, 1985). Shape and size remain relatively constant during migration, as dunes tend to establish a dynamic equilibrium with the controlling variables

(Norris, 1966; Tsoar, 1985; Watson 1987; Lancaster, 1987). Forward momentum is maintained by the continual movement of sand arriving from upwind, with subsequent erosion of the surface sand on the windward or stoss flank. Boundary layer flow compression (Jackson and Hunt, 1975: Walmsley *et al.,* 1982) supports an increased rate of sand transport towards the crest. However, significant dune lowering does not transpire, as flow separation also occurs in the crestal zone (Tsoar, 1985; Lancaster, 1987).

2.4.3 Two-Dimensional Morphology

Dune profiles in predominantly uni-directional winds are asymmetrical. A dune establishes itself above a patch of sand that began accumulating from differences in surface roughness and the sand holding abilities of the wind over the surface. Vertical growth develops a dune that has its upper form subjected to faster winds than lower morphological elements. Thus the upper components move forward, resulting in an asymmetrical shape. Shear reaches a local maximum near the crest, after which deposition must occur. Meanwhile, erosion of the windward slope, and deposition on the leeward slope, must advance the dune forward and maintain the asymmetry (Cooke *et al.,* 1993).

2.4.3.1 Windward Toe

According to Cooke *et al.,* (1993) the toe of the windward slope is a point of considerable change where many feedback adjustments occur. Slope angle and roughness are two features that greatly alter in comparison to the surrounding interdune. Tsoar (1985, 1986), Tsoar *et al.,* (1985), Howard and Walmsley (1985), Livingstone (1986) and Lancaster (1987) all reported a decrease in wind velocity at the base of low, isolated dunes. Such reduced wind-speed subsequently diminishes the sand carrying capacity of the wind, thereby leading to accumulative deposition of sediment at this zone (Cooke *et al.,* 1993). Although mathematical modelling predicts the phenomenon (Howard *et al.,* 1978, Jensen and Zeman, 1990), the reported absence of sediment piles at the bases of desert dunes may be explained as the adjustment to the curve of the windward slope so that shear and sand transport can be maintained (Wiggs, 1992).

2.4.3.2 Stoss Slope

The windward slope is a component that is in continuous feedback relationships with slope angle, surface roughness, wind flow and sand transport (Cooke *et al.*, 1993). Upslope increase in shear velocity is vital if sand eroded from the windward slope is to remain in transport. The form most conducive to this condition is a straight slope at a low angle. Hastenrath (1967) identified the dunes of the Arequipa region of Peru, to have stoss slopes that are straight, with low angles between 5° and 10°.

Modelling by Howard and Walmsley (1985), Walsley and Howard (1985) and field studies by Reid (1985) and Warren (1988b) have shown that erosion and deposition are sensitive to minor variations in the shape of the windward slope, and to the value of roughness height. However, the controls on windward slope form are obscure, with the roles of grain-size, three-dimensional shape and slope angle considered to be of major importance (Cooke *et al.*, 1993). Controversy exists for windward slope models, as Lancaster (1985a) proposed that high winds generate steep slopes, whilst Howard *et al.*, (1978) and Wipperman and Gross (1986) suggested that high wind velocities produce gentle slopes.

2.4.3.4 Crest and Brink

Dune crests have two end-member forms:

- straight to the brink, where crest and brink are coincident
- a separated crest and brink, where a broad domed convex shape occurs (Cooke *et al.*, 1993).

The extreme of crest-brink separation is shown by the dome dune form. Various workers (Bagnold, 1941; Walmsley and Howard, 1985; Lancaster, 1985a and Hesp *et al.*, 1989) have suggested that the interaction and small variation between wind velocity and direction, roughness height, and the shape and intensity of the lee separation bubble, operate as a complex feedback system. Cooke *et al.*, (1993) suggested that the lee slope to be more sensitive to minor parameter fluctuation than the windward slope.

Five fundamental explanations for differences in crestal configuration exist. These are:

1. Hastenrath, (1967); McKee and Moia, (1975); Lancaster, (1985a, 1987) envisage dune development to be sequential evolution. Dome dunes are perceived as the precursors of dunes with straight slopes. However, as McKee (1966) pointed out, dome dunes can be mature or equilibrated meso-forms juxtaposed to dunes with slipfaces, such as the gypsiferous White Sands dunes of New Mexico. Furthermore, Fryberger *et al.*, (1984) have reported dome dunes of all sizes in the Jafurah sand sea (Saudi Arabia) and the reverse evolutionary sequence of barchan to dome is not unrecorded (Cooke *et al.*, 1993).

2. For the crest of a dune to build higher than the brink, the wind would need to decelerate upwind of the crest, to allow deposition to occur. Mulligan (1987) recorded wind shear to peak before the crest. However, Lancaster (1987) has observed shear to be at a maximum at the crest. As Cooke *et al.*, (1993) explained, and the model of Jensen and Zeman (1990) showed, these apparently contrary and ambiguous results may record different stages of dune development, i.e. when a dune is in a state of non-equilibrium. Aerodynamic development of a separated crest-brink has been demonstrated by Jensen and Zeman (1990), who modelled a unidirectional wind undergoing deceleration prior to reaching the crest. Deposition of sediment increased the height of this zone in preference to the region forward of the deceleration zone. This lead to a convex separation between crest and brink. By maintaining the shear velocity to the crestline, a straight slope was sustained.

3. With varying wind speeds the point of maximum wind shear would be in a state of constant migration (Lancaster, 1987; Hesp *et al.*, 1989). This would therefore lead to a migratory point of maximum deposition. Crest-brink separation would persist only until a new wind had achieved a quasi-equilibrium where crest and brink had become coincident. Lancaster (1985a) has reported that in light winds the lower element of the windward slope can be inactive, yet the crest can be under transport by as much as 1 : 58 comparatively. Light winds are therefore capable of lowering a crest, while the base is relatively inactive, so creating a convex profile, if not crest-brink separation (Cooke *et al.*, 1993).

4. Seasonal changes in wind direction have been reported to cause crest-brink separation (Havholm and Kokurek, 1988; McKee, 1966; 1982; Burkinshaw *et al.*, 1993).

5. Both Bagnold (1941) and Hastenrath (1967) suggested that the crest-brink separation may be a consequence of lag between changes in shear velocity and the rate of sand transport, and would depend upon sand grain size and wind speed. This would shift the point of maximum sand transport downwind of the point of maximum shear stress, the crest. The point where deposition took over from erosion, the brink, would be even further downwind. This delay would be progressively less marked as the dunes grew (Cook *et al.*, 1993).

Conclusively, it is evident that no single hypothesis of crest-brink development adequately explains the global and laboratory observations for crescentic dunes. As Cooke *et al.*, (1993, p. 333) state, "More than one [process-model], may be in operation at any one site, and any one time. However, few of the models have been adequately tested."

2.4.3.5 Lee Slope: Flow Separation and the Lee Eddy

As a dune grows vertically, there is a point at which the wind can no longer follow its form, and flow separates from the bed. The detachment point is achieved at lower lee-slope angles in winds of high turbulence. The separated flow reattaches downwind at a point that is determined by the height of the dune and velocity and turbulence of the wind. From the reattachment point, the wind flows both back towards the dune, forming a separation bubble, and onwards down wind (Cook *et al.*, 1993). Hoyt (1960) suggested that in the lee of dunes, ripple patterns indicate that in high ambient wind speeds, wind velocities are great enough to transport finer sediments back towards the slipface. Significantly, these eddies can bring light plant debris to the slipface, and keep it there (Anderson and Buttiker, 1988 *in* Cooke *et al.*, 1993). Monod (1958, *in* Cooke *et al.*, 1993) recorded yardangs and evidence of excavation into the soft sands between transverse dunes in eastern Mauritania, while leeward wind erosion was noted by Sharp (1979) and Sweet *et al.*, (1988) for the Algodones dunes of southern California, and on the Oregon coast (Hunter and Richmond, 1988).

Erosion, hence morphological alteration of the slipface by lee eddies, has not been recorded by any observers (Cooke *et al.*, 1993). Sand is apparently not moved upwards, even at high wind velocities. Although, a contrary viewpoint is offered for the Kelso dunes, California (Greeley and Iversen, 1985, p. 206). Nevertheless, aprons of fine sand are commonly found at the base of the slipface (Cooper 1958, 1967; Hoyt 1966).

2.4.3.6 Lee Slope: Grainfall and Avalanching

Grainfall on the lee slope begins at varying lee angles and ambient wind conditions. Kocurek *et al.*, (1990) have observed grainfall to commence at an angle as low as 10°. Anderson (1988) found a bulge downwind of the brink, which he attributed to the slipface falling away at an angle greater than the mean return angle of grains, and to drag on the falling particles when they fall into the quieter conditions in the lee of the brink. The bulge position according to the Anderson model would depend on shear velocity and grain-size, where coarse grains with low lift velocities would fall near the top of the surface and finer ones further out. However, this contrasts with the model of Chakrabarti and Lowe (1981) in which coarse grains were carried further (Cooke *et al.*, 1993).

Avalanching (sand flow) on the lee slope occurs when grainfall builds up the upper slope beyond a critical angle, causing sand flows down the slipface. The point of slope failure due to the overloading by grainfall is termed the pivot point, where the lee slope is reduced from an angle of initial yield to the angle of repose (residual angle after shearing). According to Bagnold (1966), Allen (1970) and Carrigy (1970) the difference between the angles of initial yield and repose is *c.* 2.5°. However, the reported residual angle of most of the slipface varies slightly, e.g. 30° to 32° (Carrigy, 1970; Allen, 1971a), 33° (Chepil, 1959), 34° (Fryberger and Schenk, 1981), (Finkel, 1959; Hastenrath, 1967; 1978) 33°±1° (Inman *et. al.*, 1966) 31°- 34° (McKee, 1979c). From the pivot point to the brink, a scarp cuts back upslope, feeding a sand flow avalanche. The flow narrows or "bottle necks" below the pivot point, and then rapidly expands downslope until it reaches a constant width. The velocity profile within the flowing layer is parabolic, with a surficial low velocity layer, a fast flowing centre and a rapidly declining velocity towards the base (Cook *et al.,* 1993). The sand flow tongues tend to be approximately the same width across the slope, are only a few centimetres thick (Lowe 1976) and move at about 0.2 ms^{-1} (Hunter 1985). Subsidiary sand flows may develop on either side of the main tongue (Hunter 1977b). Avalanche flow is turbulent, mixing the grains, and this in itself must dilate the moving mass of sand (Cooke *et al.,* 1993).

2.4.4 Three-Dimensional Morphology

2.4.4.1 Age, Form and Process

A barchan dune is a member of the crescentic family that is isolated on a firm coherent basement (Cooke *et al.,* 1993). Small barchans (<1 m high) may have 'memories' that last only a few hours, or for those c. 1 m high seasonal change of winds can bring about morphologic change (ephemeral dunes). Meso-dunes (3-10 m high) can, however, have 'memories' that are as long as 1-30 years, although such types are rare and confined to a directionally constant annual wind regime (Bagnold, 1941; Finkel, 1959; Sharp, 1964; Cooke *et al.,* 1993). Such form 'memorise' the ambient annual wind regime by obtaining significant size. On the other hand, mega-dunes are those on which meso-dunes are superimposed (draa) and are considered palaeo-features of geologic age (Wilson, 1972c).

The development of the barchan (crescentic) form is brought about when wind speeds are greater close to a coherent surface, than wind speeds over dunes. Saltation is impeded by self interaction and collision between particles, thus retarding gross dune motion, while the higher desert surface winds sweep the dune horns (wings) forward. Some of the sand arriving from upwind is channelled around the flanks and leaves from the horns, moving downwind, where it is shepherded into a well defined set of vortices (Cooke *et al.* 1993).

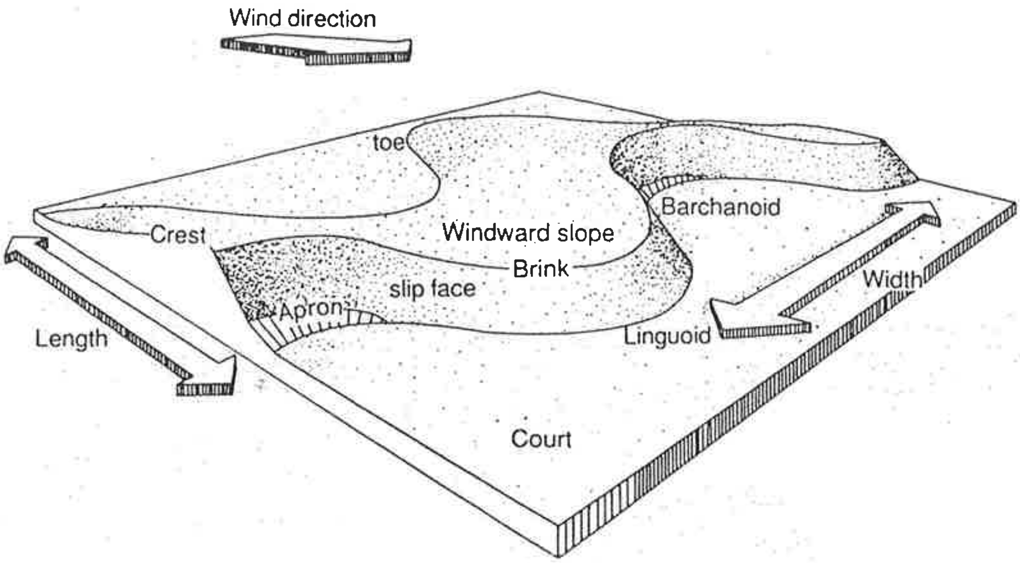

Figure 2.6 Characteristic morphological elements of a crescentic dune form evolving under the influence of a uni-directional wind. (from Cooke *et al.*, 1993)

The barchan 'court' is kept partially clear of accumulating sand by the turbulent wake in the lee of the slipface (Knott 1979). Both a horizontal vortex shed from the brink, and vertical vortices shed from the flanks, cause this turbulence (Coursin, 1964). The extent of this sand sweeping process is shown by dunes that are only 10 m high having sand-free courts as long as 5 km (Cooke *et al.* 1993).

2.4.4.2 Barchanoid and Transverse Dune Form and Process

The extensive crescentic form of transverse dunes shows variation in size that can have both length and width of a few metres to many kilometres. Transverse dunes are formed in the same dynamic environment as are barchans but are fed by a more abundant sand supply. The two types of dune (barchan and transverse) differ in that the barchanoid sections are joined by linguoidal links (Fig. 2.6) that are usually lower and sometimes coarser grained (Cooke *et al.,* 1993). As observed by Araya-Vergara (1987) barchans can evolve to transverse dunes and transverse dunes to barchans where variability exists between sand supply and wind direction. Barchans can reappear from a transverse form if sand supply diminishes and/or wind speed increases (Cooke *et al.,* 1993).

2.4.4.3 Integrated Transverse and Linear Dune Form and Process

Examination of aerial photographs of the linear dunes of the Finke River, N.T., shows the morphology of Mabbut and Woodings' (1983) network and Type I[2] linear morphologies to be very similar to the seif-linear form identified by Bagnold (1941) and Tsoar (1984). The development of nascent Type I dunes originating at or near the channel, or on the lee margin of a sandy floodplain, generates a train of short transverse crests from where the barchanoid or horn elements intersect with the stoss of another. In this manner, a longitudinal ridge or train of transverse components is formed. According to Mabbut and Wooding (1983) the dimensions of this longitudinal structure are only slightly smaller than

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[2] Type I dune pattern shows a spacing in the range of 200 - 400 m, a high interconnectivity between dunes, marked asymmetry and sandy swales (interdunes). Type II dunes are spaced in the range of 400 - 800 m, show moderate to low connectivity and asymmetry, with the swales floored by loamy alluvium. Type III dunes have spacing > 800 m, consist of massive dune ridges separated by stony plains with small playas. (Mabbutt & Wooding, 1983, 51 - 69).

those of the Type I pattern, but lack the regularity and continuity of the Type I pattern. Crescentic links between the transverse trains are also common, and have been termed *network* dunes. As the distance from the river increases, the spacing and dimensions of the transverse dunes become more like Type I, although the crescentic link between longitudinal ridges remains. According to Mabbut and Wooding (1983), the Type I (seif-linear morphology) pattern originates close to the area of fluvial input and/or the current aeolian/fluvial reworking of older sands. At these localities the substrate of the area of dune initiation, and the substrate of the developing and adjacent interdune, show a minimal contrast to the sand-transporting winds (Mabbut and Wooding, 1983). However, in areas of relatively consolidated substrate, such as a stony or gravel surface, dunes of increasing organisation (Type II and Type III) are the prevalent morphologies.

Both organisation and spacing increases downwind of Type I, with each type progressively forming upon a more consolidated desert floor. Type III dunes build in areas where the greatest contrast occurs between the surface texture of the dune and interdune corridor (Mabbut and Wooding, 1983), which has suggested an association between dune type, dune spacing and substrate texture. The alliance of a firm desert pavement (alluvial gravels, gibber plain or low tablelands) with dune building allows the derivation of linear morphology that demonstrates the coalescence and downwind termination of dune segments, and an increased dune spacing. Thus each type of dune pattern corresponds with a different physiographic setting. Besides wind direction and intensity, and sand supply, dune morphology is significantly influenced by the nature of the desert substrate upon which dune initiation occurs.

2.5 Linear Dunes

2.5.1 Morphology and Morphometry

Linear dunes, also commonly referred to as seif or longitudinal dunes, are one of the most common terrestrial dune-forms. Landsat imagery has revealed that an estimated 30% of all depositional aeolian landforms on Earth, are linear dunes (Lancaster, 1982a). From the orbital platform, linear dunes are seen to be the dominant morphology of terrestrial desert ergs, excepting the Erg Oriental in the Sahara and the Nafud and Jafura in Arabia. Many linear dunes are simple forms, occurring in extensive dunefields, with individual dunes usually about 20 m high and up to 1 km apart (Goudie, 1970; Twidale, 1972), though compound and complex varieties as in the Namib can attain heights of up to 200 m (Lancaster, 1982a; 1982b) and exist in a variety of wind regimes (Fryberger, 1979).

A great deal of morphological variation exists within the global population, which consequently evokes a variety of theories of origin and evolution for linear dunes. As will be shown, many of these hypotheses are erroneous in nature. However, it is also evident that no single hypothesis can sufficiently account for the initiation, growth, maintenance, parallelism, obliquity or form of all linear dunes in all deserts. The variables of wind regime, sand supply and climatic setting are many and varied for the environments in which linear dunes occur, and as inferred by Cooke *et al.*, (1993), composite rather than individual explanations (models) may more than likely be the requirement. Unlike crescentic forms, the genetic link between all linear dune types is not successive.

Early investigators such as Price (1950) used the observations of Bagnold (1941), Capot-Rey (1945) and Madigan (1946) to recognise three subtypes of linear dune based upon the formative parameters of wind regime, sand supply and vegetative cover. These classes were:

- the Australian sand ridge, wholly or partially vegetated, smoothly rounded and asymmetrical in profile
- the broad, bare high ridge with sparsely vegetated plinths and, several lines of slip faces aligned longitudinally to the dune (also termed draa)
- the bare ridge with sinuous crestline and numerous re-entrant barchan-like slipfaces, commonly termed the seif dune.

However, in a more recent classification McKee (1979b) subdivided linear dunes into simple, compound and complex. The morphology of a simple linear dune was shown by a single narrow dune ridge with a straight or sinuous crestline of rounded or sharp profile. Secondary dune development was absent. Further to this, two subtypes of simple linear dune are commonly recognised and are distinguished by the classification of Breed and Grow (1979). Subtype One is recognised by a relatively short, sinuous, sharp crested dune with a pointed downwind end. This dune is commonly referred to as seif, as encountered by Bagnold (1941) in the Libyan desert and Tsoar (1974, 1978) in the Sinai. Subtype Two is characterised by long, straight dunes as typified by those of the Kalahari and Simpson deserts. A classification by McKee (1979b), Lancaster (1982a) and Tsoar (1989) also further subdivide simple linear dunes into three types:

- lee dunes
- vegetated-linear dunes
- seif linear dunes

noting that all three types differ in their genesis and evolution, although they may appear morphologically similar.

Lee dunes form in the lee side of an obstacle such as a cliff, boulder or bush while their size is proportional to that of the obstacle (Tsoar, 1989). Vegetated-linear dunes on the other hand are of low and rounded profile, having a varying degree of vegetative cover. Dune stabilisation accompanies vegetative encroachment. These dunes also display an exclusive attribute known as Y-junctions, where two vegetated-linear dunes converge into a single ridge. Such dunes are typical of the central Australian deserts (Madigan, 1936, 1946; King 1960, Mabbut 1968, Clarke and Priestley, 1970; Twidale, 1972a, 1972b, 1980, 1982; Wasson, 1983b) and of the Kalahari in South Africa, Arizona and California as well as the Indian and Negev deserts (Tsoar, 1989). The common factors of all these schemes of classification is that linear dune morphology is a function of:

- the wind regime.
- sand supply.

- vegetative stabilisation and obstruction to the wind regime and sand supply.

Compound linear dunes consist of two or more closely spaced parallel or converging narrow dune ridges on the crest of a much wider and larger plinth. Such structures have also been termed anastomosing complexes by Holm (1960). The third category of complex linear dunes are characterised by being larger than either the simple or compound varieties. These dunes reach heights of 100-150 m in the Namib and between 150-200 m in the Arabian deserts. Both localities show a single main ridge, with regularly spaced, sometimes star-form, peaks en-echelon with a sharp sinuous crestline joining them, and a major slip face to one side. Secondary dunes, often oblique or transverse to the main trend, are frequently developed on the plinths (Lancaster, 1982a).

Furthermore, Hunter *et al.*, (1983) classified linear dunes according to the relationship between dune trend and the long term resultant sand transport direction. In this scheme of things, linear dunes lie roughly parallel to the resultant sand transport direction (<15° angle), while oblique dunes stretch at angles between 15°-75° to the resultant sand transport direction.

However, there are deficiencies in the schemes of Hunter *et al.*, (1983), whose classification is based upon morphodynamics alone, and also that of Lancaster (1982a) who bases his observations on McKee's (1979b) scheme of classification. Both lack the ability to distinguish variants of dune form within each of the categories, as well as the inability to recognise transitional forms between the categories of simple, compound and complex dunes. All in all, these classifications are oversimplified and are undiagnostic for a global sample of morphologies. Tsoar (1989) also claims that it is incorrect to assume that all linear dune morphologies are developed in a single morphodynamic manner, criticising investigators such as Warren and Hyde (1983) and Hunter *et al.*, (1983) for making this assumption. He asserts that to place linear dunes under a single genetic classification is erroneous. As noted by McKee (1979b),"For all dune types - simple, compound and complex - an almost infinite number of varieties occurs. The dune varieties are too numerous to describe or even classify by name herein, for they probably represent transitions from one basic type to another...".[3] Thus it should be emphasised that the various schemes of categorisation for linear dunes do not account for all morphological variants, but are general attempts at classifying a large and diverse morphometric class of dune-form. At best the classes of simple, compound and complex are the most useful, yet general, categories that have been proposed.

2.5.2 Erosional Origin

From the observation of dune trend and regional wind regimes, many workers such as Madigan (1936) and Folk (1971b) have documented the alignment of linear dunes and the direction of the prevailing winds. The hypothesis of prevailing wind drift (King, 1960) states that dune development is caused by deflation of adjacent interdune areas. Essentially the dune is a yardang, with a surficial covering of mobile sands. Erosion of a pre-existing deep sediment pile has been argued by Blandford (1876) and Verstappen (1968) for the Thar Desert, Aufrere (1928) for the Sahara and Melton (1940) for Arizona. In the Sahara and Arizona, a linear morphology has developed from erosion of U-shaped dunes (Lancaster, 1982a). Examination of the Simpson (Lake Eyre, Australia) region by King (1960) and Folk (1971b), supports the idea of initial wind erosion channelling the desert floor to develop ridges of alluvium with later deposition being primarily upon the alluvial ridges. Folk (1971b, 1976) interlinked interdune erosion to the notion of roller vortices. Although King found the alluvial core to be approximately 6 m deep, Mabbut and Sullivan (1968) found that in other areas of the Simpson Desert, the sediments beneath the dunes continue uninterrupted in both areas of unconsolidated alluvial sediments and indurated gravels (Lancaster, 1982a). King (1960) also noted that the typical Y-junction of Australian linear dunes was evidence of the deflation of parabolic dunes. A similar scheme was argued for the linear dunes of the Thar desert (Verstappen 1968, 1970; Wasson *et al.,* 1983) where

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[3] McKee E. (1979) *Global Sand Seas USGS Professional Paper 1052*, p13

erosion of parabolic dune noses determines the establishment of linear morphologies. However, the converse of this was proposed by Madigan (1936) and Mabbut and Sullivan (1968) who regarded these junctures as the convergence of adjacent ridges caused by the deflection of the downwind ends of growing dunes. Mabbut and Sullivan also suggested that this interpretation aids in explaining the often asymmetrical nature of these dunes.

2.5.3 Depositional Origin

Madigan (1936, 1946) supported the idea that linear dunes develop from deposition of sand in strips aligned with a prevailing dominant wind direction, and rejected the notion of deep pre-existing sand deposits in the Simpson Desert (Lancaster, 1982a). Bagnold (1941) also described that wind moving over a gravel surface produced parallel, longitudinal, wavy ribbons of sand. When the wind speed decreased, these longitudinal sand ribbons further developed into linear dunes, as surface roughness differences have encouraged deposition on the sandy areas. Further developments of this model were made by Glennie (1970) who proposed that once dune development had occurred, pressure gradients created between dune crest and interdune corridor allowed sand laden surface eddies to spiral from the interdune corridor to deposit upon the dune. Although this model is a succinct explanation of dune development, their exists no field evidence to support this hypothesis (Lancaster, 1982a).

Bagnold (1953) developed the idea of vertical convection cells originating from thermal instability of air over desert surfaces, with still air vertical cells developing into pairs of longitudinal roller vortices spaced at a distance three times that of the convectional layer when in moving air. This helical motion of air would impart an alternate left and right hand oblique component to the surface wind, which would sweep sand into long parallel ridges spaced the width of a vortex pair apart (Lancaster, 1982a).

Hanna (1969) further developed Bagnold's principle of helical roll vortex deposition, claiming that dunes were the consequence of convergence between the surface winds of two counter-rotating helical roll vortices formed in trade winds across desert surfaces. Both laboratory and field evidence demonstrated to Hanna (1969) that helical roll vortices were of widespread occurrence. Comparisons of linear dune spacing with the expected dimensions of helical roll vortices in subtropical desert areas revealed that the vortex should have a wavelength of between 2 and 6 km (i.e. 2-3 times the convective layer thickness of 1-2 km) (Hanna, 1969). However, as pointed out by Lancaster (1982a), a major flaw occurs with this argument. If helical vortices of an approximate circular form create linear dunes, then the spacing should be at two times the diameter of the vortices, or 4-12 km apart. The evidence of Breed and Grow (1979) shows that the maximum spacing of linear dunes in regular patterns is between 3 - 3.5 km, and that in areas such as the Simpson and Kalahari deserts, dune spacings are commonly < 1 km, dimensions much smaller than any reported atmospheric helical roll vortices (Le Mone, 1973). This discrepancy between dune spacing and the meteorological origin of helical roll vortices implies that linear dunes are not the product of such atmospheric motions (Lancaster, 1982a). Furthermore, there are no observations of sand transport from interdune to dune or evidence for regional scale helical roll vortices in areas of linear dune dominance. Tsoar (1978) found that vortex flows are restricted to the separation zone immediately adjacent to the crest of the dune, while the measurement of atmospheric movements of roll vortices and their velocities (Angell *et al.*, 1968) show that horizontal velocities are between 2-6 km h^{-1}. Such motion would equate to minimal bulk sand transport and unlikely to saltate the volumes of sand necessary for dune building. Despite such fundamental problems with the helical roll vortex model, wide adoption of this hypothesis has occurred (Wilson, 1972; Cooke and Warren, 1973; Warren, 1979). Amended versions of this model combine both wind drift and helical roll vortices into a scheme which has vortices moving sediment from pre-existing deposits into uniform evenly spaced dunes (Folk, 1971a; 1971b).

Considering the fundamental inadequacies of both the wind drift and helical roll vortex theories, it is undesirable that a composite model should evolve from these ideas. To further illustrate this point, the fact that dune landscapes do exist upon bedrock and indurated substrate, where the former existence of a deep sand cover was unlikely, is evidence that the classical wind drift and helical roll vortices models are inadequate and do not satisfactorily explain the origin of linear dunes.

2.5.4 Resultant or Bi-Directional Wind Models of Origin

Bagnold (1941) observed that many linear dunes extend parallel to the resultant direction of sand movement. This is demonstrated by the manner in which movement of a barchan dune in a bi-directional wind regime is greater for the horn, that receives sand from strong oblique winds, and is extended or elongated by a gentler secondary wind direction. Cyclicity of this process would form a seif-linear dune with regularly repeated summits and a sharp sinuous crest. Barchans with extended arms are common in some parts of the Sahara (Clos-Arceduc, 1967). Warren (1976) reported that linear elements in the Nebraska sand hills are the result of bi-directionality in the wind regime, while Lancaster (1980) and Kar (1987, 1990) identified similar occurrences in the Namib and Thar deserts, respectively.

Bagnold (1941) noted that such a hypothesis is perhaps not a universal model for the evolution of all linear dune-forms. However, it does establish that linear dunes are the result of bi-directional wind regimes that extend in the resultant direction of sand movement. The diurnal reversal of slipfaces near the crest of linear dunes of the Simpson desert, offers supportive evidence of dune construction by a bi-directional wind regime with a dominant directional influence (Wopfner and Twidale, 1967; Clarke and Priestley, 1970; Twidale, 1972).

Tsoar's (1978) micro-meteorological work in the Libyan desert demonstrated that winds which blow at < 90° to the crestline are diverted to flow parallel to the dune crest. This, in effect, will elongate the dune downwind. Hence any wind from a sector of 180°, centred on the dune, will be deflected in this manner and linear dunes will form in a wind regime where the winds blow from one wide sector (Lancaster, 1982a). The diverted winds have a relatively high velocity and do not deposit sands directly on the lee flank, except in a small slip-face zone. Instead, they move along the dune and in the attachment zone they are even eroded from the lee flank. Significant deposition on the lee flank occurs only when the angle between the crest line and the wind angle approximates 90°. Most deposition on the dune is not of the avalanche type, but is rather lee side accretion (Lancaster, 1982a). Seasonal or diurnal changes in the wind direction will cause changes in the erosion/deposition patterns on the dune that can result in a sinuous pattern of narrower, lower saddles and higher, wider peaks. For winds from either side of the dune, the crest line will consist of flow parallel and flow transverse segments, with erosion occurring on the flow parallel segments and deposition on the flow transverse segments. The dune narrows when erosion exceeds deposition, while other areas are sites of nett deposition where the dune will widen and grow in height. However, as the area of deposition is spread over a greater area, the growth rate will progressively decrease, eventually establishing a dynamic equilibrium with peaks and saddles at a uniform height (Lancaster, 1982a). Tsoar (1979) suggested that the tendency will be for the dune to extend downwind, with the sinuosities gradually moving along the dune. The crestline morphology and regularity of the crestal sinuosity are dependent upon the consistency of the formative wind direction (Lancaster, 1982a). If winds on either side of the dune are of equal strength, peaks and saddles do not develop. Tsoar's model is significant in the study of linear duneforms, as he has demonstrated by relating morphology to aeolian process, the manner in which linear dunes can maintain their morphology as well as grow longitudinally.

Further comment by Lancaster (1982a) also implied that linear dunes genesis is commonly associated with wind regimes with a seasonal variation of direction over 120°-180°. Thus, it is likely that linear dunes, due to their global abundance, are subjected to wind from a wide convergent sector or two major convergent directions. Furthermore, linear dunes subjected to a single major wind direction are prone to erosion and segregation. Thus bi-directionality is necessary for a linear dune to develop, survive and evolve. Nevertheless, a persistent wind from a single direction will have a major effect on alignment of a dune, but little evidence exists to suggest that linear dunes form parallel to this wind direction. As commented by Tsoar (1978), dune alignment will tend to be at oblique angles of 20°-30° to the persistent wind directions.

One characteristic of linear dune development, not yet satisfactorily explained by any model, is the regularity of spacing between adjacent dunes. Any disturbance of the wind by dune height, is estimated to be *c.* 12-15 times the dune height in length (Tsoar, 1978; Oke, 1978). Because the winds blow obliquely to linear dunes, separation between the dunes along the wind direction is much greater, as observed in the Simpson desert by Twidale (1972). Both Twidale (1972) and Lancaster (1981b; 1981c) indicated that some form of aerodynamic control of dune spacing exists, as portrayed by the height/spacing correlations of linear dunefields.

2.5.6 Linear Dune Evolution from Barchans

2.5.6.1 Bagnold - Tsoar Horn Elongation Models

Bagnold (1941) initially proposed a model of seif-linear dune origin and evolution from the barchan form. Tsoar's (1984) work on the Tenere desert of the Sinai demonstrated the genetic development of seif-linear dunes from barchan and transverse forms, when barchans move into areas of bi-directional winds. In this scenario, horn elongation is preferential to the dominant wind sector of the regional wind regime. The observation of barchan horn elongation has been made in many arid zones and by many workers (Kerr and Nigra, 1952; Holm, 1960; Norris, 1966; Clos-Arceduc, 1967, 1969; Tsoar, 1974; Warren, 1976; Grolier *et al.*, 1976; Lancaster, 1980; Kar, 1987, 1990). Figure 2.15 depicts the Bagnold model, where a symmetric barchan formed by gentle, steady winds develops in a predominantly unidirectional wind regime. However, with storm winds that are oblique to this main direction, the elongation of a single barchan horn occurs until it encounters the sand stream moving off the second non-elongated horn. From this point of sand flow

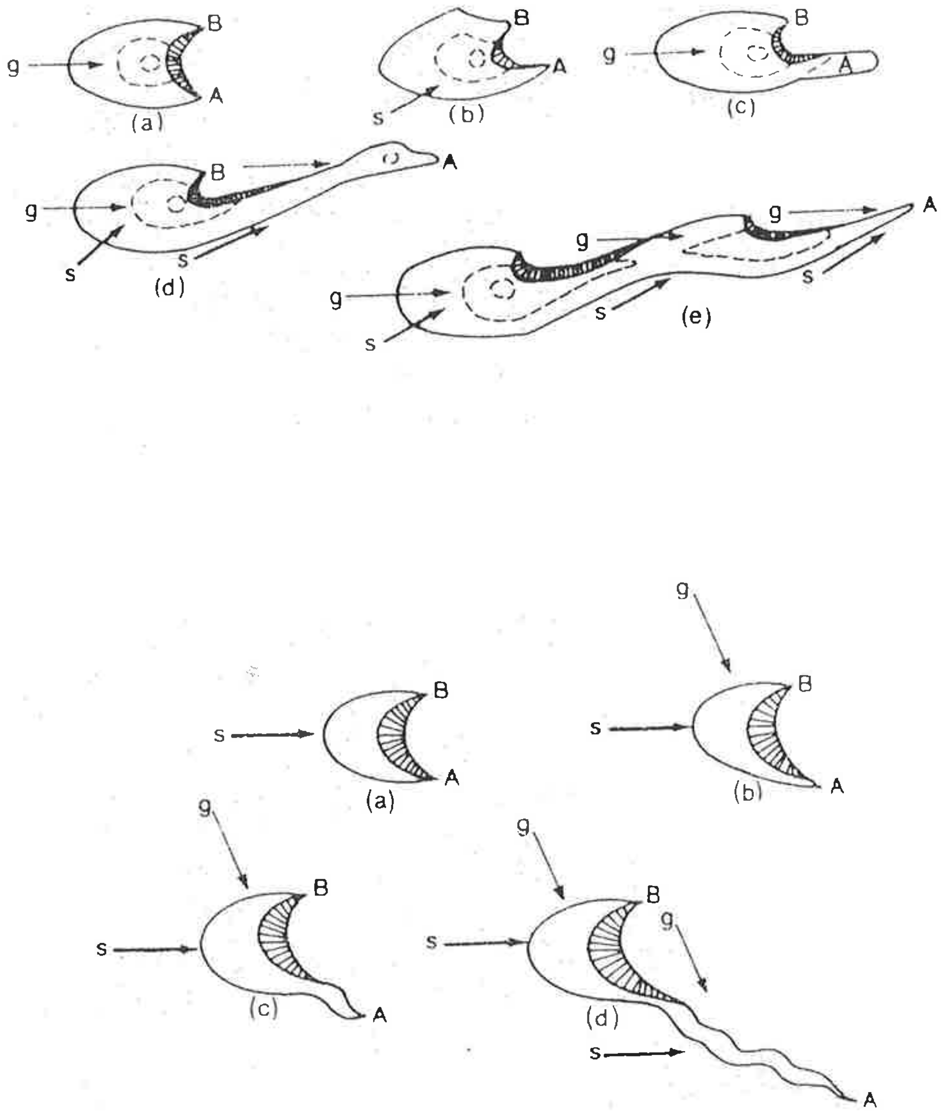


Figure 2.7 The formation of seif-linear dunes from barchans, after Bagnold (1941) (upper figure). Direction *g* is the primary wind vector, while *s* represents storm winds which elongate horn A of the barchan. An amended model of seif-linear evolution by Tsoar (1984), (lower figure), where two main directions blow obliquely on two sides of the barchan. In this instance, horn A is elongated with seif-linear development being realised by deflection of the two winds parallel to the crestline of the horn extension.

mergence, the horn will grow rapidly, fed by winds from both directions, and turning into a seif-linear form. The Bagnold model shows elongation of the horn only on the side exposed to the storm wind, and that the angle between the orientation of the seif dune and the barchan ridge line is *c.* 80°(Fig. 2.7).

An amended model of the Bagnold (1941) hypothesis, has established evidence for the development of the seif-linear by oblique winds either side of the dune (Tsoar, 1978, 1982, 1983a, 1984). Each wind direction is deflected on the lee side of the dune and flows parallel to the crestline along that side. Strong winds are necessary to bring about sand accretion on a desert surface (Bagnold, 1941), so that barchans are initially oriented to the direction of strongest wind (Fig. 2.7). After build-up of an initial barchan, a gentler oblique wind direction affects the dune's morphology, highlighting the importance of a bi-directional wind regime in developing seif-linear morphology. Horn *(a)* is eroded by gentle secondary winds, while horn *(b)* on the opposite side of the secondary wind, is located so that two wind directions encounter it from the two sides which can develop movement of sand along both lee sides of the horn, and elongate it to form a seif-linear dune (Fig. 2.7). The main differences between the Bagnold and Tsoar models are:

- the Tsoar model shows the barchan forming from the strongest winds, with the horn being elongated by gentler secondary winds;
- Tsoar shows that only the horn opposite the secondary wind is capable of elongation;
- the angle between the orientation of the barchan ridge line and the alignment of the seif dune is 120° in the Tsoar model rather than 080° of Bagnold;
- both winds bring about elongation of the horn.

Lancaster (1980) provided support for the original Bagnold model. From his studies of the Namib desert, he showed that the development of seif-linear dunes in all deserts may be explained by variations on the Bagnold theme.

2.5.6.2 Kar Model of Barchanoid Coalescence

A model proposed by Kar (1987, 1990) where longitudinal (linear) dunes in Jasalmer district of Rajasthan India, and barchanoid and parabolic morphologies are revealed to be transitional and genetically linked. Kar (1987) subdivided the linear dune forms of this region into four classes:

- Simple unidirectional - no barchanoid crest
- Complex unidirectional - barchanoid crest
- Simple bi-directional - no barchanoid crest
- Complex bi-directional - barchanoid crest

The simple forms are typically stable, without any indication of reactivation, while the complex varieties show a barchanoid crest developing upon a once stable linear dune. Observations confirm that where the linear dunes are now forming, they are formed by the coalescence of narrow belts of barchanoids along the crestal zone of linear dunes. These barchanoid crested linear dunes therefore qualify to be classified as complex linear forms within McKee's (1979b) classification. Similar complex longitudinal (linear) dunes have been reported in the Sahara (Mainguet, 1984) and the Taklomakan desert (Zhu Zhenda, 1984).

In all reported cases of evolving linear dunes, a stream of closely spaced barchanoids in narrow chains with corridors of sandy flat were observed. The actual sequence of events from the formation of barchans to linear dunes can be tentatively reconstructed on the basis of observing a wide range of linear dunes in different stages of development (Kar, 1987). Kar's evolutionary model for the development of linear dunes from barchan dunes follows the procedure:

- formation of isolated barchans over a plain surface;
- gradual colonisation of a long strip by closely spaced barchans which interlink;
- with increased sand supply, height of the interlinked barchans increases;
- a critical height is reached when increased shear on the windward slope of the crestal segments slows down crestal growth;

- leeward airflow separation creates smaller vortices on the lee slope, the length of which is proportional to the ridge height;
- the eventual formation of small transverse basins with secondary vortices at the micro-level, are topped by high flow lines;
- sand saltating forward from the crest and from slipface avalanche is trapped by the reverse flow of these local vortices and begins accumulating at the base of the lee slopes as small cones;
- some of the dunes developing a low elongated ridge in the direction of the wind between the two summits of a barchanoid ridge;
- the elongated ridge assists in the deflection of minor eddies towards the central axis of the slipface;
- the convergence of minor eddies combined with reverse flow from the top of the dune, may significantly decelerate the winds and cause deposition along the line of convergence;
- the apex of each depositional cone aligns itself roughly with the central part of the slipface;
- the dunes do not gain significantly in height from this stage onwards;
- upon reaching the crest of the barchanoid, saltating sand is trapped in the secondary flow and deposited along the leeward cone that gradually increases cone height (at the cost of the slipface) and length;
- ultimately long ridges are formed, joining dune slipface with the rear of other barchanoids;
- the transverse micro-basins between the barchanoid chains are subsequently replaced by the longitudinal micro-basins in-between the ridges, thereby shifting the secondary flow to the sides and changing basin geometry;
- infilling of the intervening basins reduces the height of the slipfaces, with the whole complex taking the shape of a linear dune;

- as the plinth of the dune gains height, and the amplitude of relief in the wavy crestal region is reduced through progressive burial, a disequilibrium is reached and the process of barchanoid formation gains momentum;
- gradually dune height increases, but the width may remain almost constant, as this is predetermined by the spacing of the vortices (Kar, 1987).

This evolutionary concept is considerably different from the models of Bagnold (1941) and Tsoar (1984), but again shows the possibility of barchan dunes being transformed into a variant of the linear dune.

2.6 Dune Sediments and Sedimentology

The sedimentological distribution of a dune is a response to the processes of sand differentiation over a dune and hence the sediment budget of erosion, transport and deposition. According to Bagnold and Barndorff-Nielson (1980), the smaller grains are more readily deflated and deposit upon a saltating surface. Consequently, an erosional surface is overloaded with coarse grains and depositional surfaces are typically abundant with fine grains. It is this process of granulometric sorting that identifies the spatial and temporal sedimentological signature of the sediment budget.

Cooke *et al.*, (1993) pointed out that many factors are responsible for the sorting process and include:

- sorting at the grain scale; shaking induced by bombardment of saltating grains may instigate removal of grains by adding to the quantity of saltation, or by the vertical settling of small grains between the inter-particle voids of larger grains that are only capable of creep movement;
- sorting on single dune slopes by wind threshold velocity increase over the windward slope towards the crest, inducing the entrainment of fine sands under the influence of higher wind speeds;

- sorting caused by the variation of wind speed and direction with seasonal or diurnal change, and hence the adjustment of slope gradient, especially at the higher levels of the dunes that are most vulnerable to wind changes;
- sorting by grainfall and avalanching of leeward slopes are influenced by slope gradient, gravity, moisture content of the sands, and the length of saltation paths;
- the accumulation of wind-borne dust, assisting in the 'smoothing' of the sediment-airflow interface, which reduces surface roughness and affects sorting;
- sorting caused by dune migration suggests that leeward deposits are returned to the base of the windward slope and that three-dimensional sorting occurs around a dune.

Sarre and Chancey (1990) have also identified the process of vertical grain-size differentiation. Working in the Jafurah and Dahna sand seas of Saudi Arabia, they observed that entire dune surfaces were consistently coarser and better sorted than sand at a depth of a few centimetres. Although coarse lag deposits can form by direct differential aeolian sorting (winnowing), the process of size segregation is not a direct interaction between wind and sand. Size segregation involves the transfer of impact energies of the saltating grains into the surface layers. This causes compressional rebound of the subsurface and its emplacement as the differentiated surface layer. The upward movement of larger grains occurs at the expense of smaller sizes, developing a coarse grained surface layer.

Aeolian sorting involves many processes. In as much as individual mechanisms do occur, each process is interactive with each other, and is continually adjusting to temporal change in the aeolian regime (Cooke *et al.*, 1993). With all other things being equal, it is most plausible that the granulometric disparity often cited for different geographical locations reflects differences in the intensity and variability of the aeolian regime and its periodicity.

2.6.1 Aeolian Grain-size Distributions

Statistical examination of the grain-size distributions of sand over various dune types has shown some interestingly controversial results for individual deserts. Many investigators (Bagnold, 1941; Alimen, 1953; McKee and Tibbitts, 1964; Glennie, 1970; Lancaster, 1985a, 1985b, 1986, 1987, 1989) have noted changes in grain-size and sorting over dune profiles. Lancaster's (1987) work showed that the crestal sands of Algerian and Namibian linear dunes are finer compared to those in the plinth. This contrasts with the observations of the Australian Simpson Desert linear dunes, where crestal sands are coarser and better sorted than those from the plinth (Folk, 1971b). To add to this ambiguity, Warren (1972) reported no change in grain size or sorting across linear dunes. Folk (1971b) suggested that poor sorting is due to the proximity of the fine-grained alluvial sand source for the dunes, and that with time, the preferential aeolian removal of finer material would develop coarser, well-sorted crests. Lancaster (1982a; 1982b) observed that dune flanks are intermediate in size and sorting, trapping weakly saltating coarse sand and receiving deposits of fine avalanche sand from the slipfaces.

Lancaster (1981a; 1981c; 1982b) eloquently explained how fine well-sorted crestal sands, and coarser less-well-sorted sands on the flanks of the complex linear dunes of the Namib, are related to the pattern of sand movement on the dunes. As sand moves onto the windward slope, coarser traction load slows with slope grade thereby diminishing the efficiency of saltation. The traction load approaching the crestal region is steadily left behind, causing a progressive fining of sand towards the crest. Slip-face avalanching develops a preferential downslope movement of coarse grains that accumulate at the base of the dune. Seasonal variation in slipface orientation causes both flanks to have deposits of coarse sands.

Few workers have recorded coarse sand on the crests of crescentic dunes, with most crestal coarseness being reported from linear forms (Sharp, 1966; Folk, 1971b; Chaudru and Khan, 1981; Barndorff-Neilson *et al.*, 1982; Lancaster, 1986; McAuthur, 1987). This may be explained by dune sand having been derived from interdune corridors. A high fraction of small mean grain-size particles, silts and clays, are more readily winnowed from the dune sands to leave behind coarser grades. This mechanism was proposed by Folk (1971b) for the granulometric character of linear dunes in the Simpson Desert. Another feasible explanation for coarser crests is the winnowing out of finer sediments from the upper slopes by high velocity winds. The residual deposit consisting of coarser particles. However, as reviewed by Cooke *et al.*, (1993) and found by Alimen (1953), Verlaque (1958), McKee and Tibbetts (1964), Glennie (1970), Embabi (1970/71), Lancaster (1981c; 1986), Sneh and Weissbrod (1983), Binda (1983), Vincent (1984), Ashour (1985) Watson, (1986b), Lancaster *et al.*, (1987), Livingstone (1987b), Hartmann and Christiansen (1988) and Tsoar (1990), it is more common for crests to be finer, better sorted and less fine skewed than the lower windward slopes. Coarse sand occurs commonly at the base of the windward slope often forming mega-ripples (Finkel, 1959; Hastenrath, 1967; Lancaster, 1981c; 1986; Vincent, 1984; Watson, 1986; Livingstone, 1987b), and on the horns of barchan dunes and linguoid of continuous transverse forms (Cooke *et al.*, 1993). Hastenrath (1967) proposed that this may be caused by the inability of coarse, windward sands to climb the windward slope of barchans, so resulting in the channelling of sediment around the curved horns of the dune.

In general, upslope fining of sand has been identified and related to a slope threshold effect by Vincent (1984), Watson (1986) and Livingstone (1987b), while Sneh and Weissbrod (1983) attributed upslope fining to sorting on the slipface, leaving only fine particles to be circulated in the upper dune elements during dune advance. As stated by Cooke *et al.*, (1993) this latter hypothesis could only continue if coarse material were able to mount the dunes for recirculation by dune advancement.

2.6.2 Aeolian Sand Morphology

Khalaf and Gharib (1985) and Thomas (1984, 1987) have shown that aeolian sand-grain morphology is notably subrounded to subangular, mean roundness increases with grain size and that roundness is inherited from source sediments (Folk, 1978; Goudie and Watson, 1981). As indicated by Thomas (1987) the highest mean roundness values do not correspond with the coarsest grade sediments, but are found within discrete grain size limits. Thomas (1987) has suggested that grain rounding in aeolian sands results from abrasion during transportation. The lack of particle roundness for larger grain sizes can be attributed to:

- the lesser mobility of these grains by the mechanism of creep rather than by saltation or suspension. Grains moved by creep or traction are probably subjected to less grain-upon-grain impact, creating fewer opportunities for edge abrasion;
- the likelihood of the largest grains being subjected to abrasion is decreased by them being mobilised less often;
- a contrast to the rate of abrasion inflicted on the largest particles is the decreased rate of abrasion, also evident on smaller grain sizes outside the limits of maximum abrasion (Thomas, 1987).

From these observations it can be inferred that the roundness of aeolian sands is only partly a function of wind velocities and frequencies during sand mobilisation (Thomas, 1987). As the parameters of wind velocity and frequency vary in individual desert settings (Fryberger, 1979), the roundness of sands from different localities will also vary. Kuen (1960) also noted that abrasion is a function of both grain size and wind velocity.

Khalaf and Gharib (1985) and Folk (1978) found that aeolian sands in Kuwait and the Simpson deserts respectively, inherited roundness characteristics from their parent materials, suggesting that insufficient abrasion had occurred for the sediment to achieve any higher degree of particle rounding. This in turn argues against possible long distance transport for the sedimentary materials. Goudie and Watson (1981) reached a similar conclusion regarding the relatively low degree of mean roundness for sands of the Namib desert dunes. However, the history of aridity for this desert is long and uninterrupted (Ward *et al.*, 1983) and offers the potential of almost continued reworking, mobilisation and transport. This ambiguity is not simply explained.

It is possible that particle roundness may be related to the size and type of dune morphology (Thomas 1987). It is suggested by Thomas (1987) that the smaller and more mobile dune morphologies, cause greater edge abrasion of dune sands thereby leading to increased roundness. Linear dunes may therefore indicate less sand mobility than do barchan style dunes, where sand reworking for individual grains for the barchan is more uniform during its ongoing advance.

2.6.3 Aeolian Ripple Morphology

Wind ripples are ubiquitous in desert environments, and are absent only where:

- very coarse, usually bimodal sand, is deposited;
- there are extremely high threshold friction velocities;
- grainfall into local areas of low wind velocity occurs;
- there are actively avalanching slipfaces (Cooke *et al.*, 1993).

Both the planimetric patterns and the orientation of ripples on dunes, indicate antecedent wind conditions and the micro-complexity of wind currents moving over a topographically varied ground surface (Sharp, 1963). Sharp (1963) clearly identified the reversal of opposed winds in the Kelso Dune complex for different times of the year via the orientation of ripple lee slopes, as well as air flow divergence over a dune due to dune shape and the existence of pressure gradients set up by differences in wind velocity as created by local topographic configurations.

Equilibrium of form can be reached, usually within minutes (Sharp, 1963; Seppälä and Lindé, 1978). Thus ripple lives are generally short, and modification of wind-ripple shape under the influence of gentle winds can also be rapid, once the 'parent' form has developed. In general, wavelength range is between a few centimetres to tens of metres, while ripple height can be less than 1 cm and up to 30 cm (Cooke *et al.*, 1993). The wavelength:height ratio ($RI = \lambda/h$) or ripple index is often referred to as being greater for aeolian ripples than for sub-aqueous forms (Tanner, 1967; Tucker, 1991). Tanner (1967) suggested that a $RI \geq 17$ attests an aeolian or swash origin. After comprehensive sampling and statistical testing, < 2% of many hundreds of measurements have been on the "wrong-side of the criterion" used to distinguish wind ripples from sub-aqueous forms. Nevertheless, Sharp (1963) and Seppälä and Lindé (1978) did not consider the ripple index to be diagnostic of an aeolian origin, as the range of indices is large (as experimentally produced by Bagnold, 1937a, $10 > RI < 70$,), and straddles the indices derived for sub-aqueous forms. However, in defence of Tanner's work, he does explain that, "ripple marks having flat tops, filled troughs, or other deformities which would affect the index, cannot be interpreted in this way" (Tanner 1967, p. 95). Obviously such discord between the results of these workers may be resolved through standardised methods of comparative analysis, that show both the raw data and the technique. Significantly, Sharp (1963) determined that the ripple index varies inversely with grain size and directly with wind velocity.

Seppälä and Lindé (1978), using wind tunnel observations, found that both mean wavelength and height increase with wind velocity. A Pearson correlation coefficient of $r = 0.998$, demonstrated that mean wavelength is highly correlated with shear velocity. As u_* increases, the range of wavelengths also increases. At high velocities, however, many small, secondary ripples can cover larger forms (Seppälä and Lindé, 1978). This relationship with shear velocity is well illustrated on dune crests in contrast to the lower morphologic localities, where greater wavelengths correspond with the increase in flow velocity that occurs over dune crests (Hunter and Richmond, 1988). Generally, ripples are flow transverse. However, ripple movement on slopes is assisted by gravity, resulting in the ripple crestline being oblique to the direction of air flow (Howard, 1977). Plan morphology varies from 2D symmetry and straight to 3D asymmetry with varying degrees of sinuosity. Sinuosity is related to both wind speed and grain-size, where at a constant shear stress, ripples composed of coarse sand are more sinuous than those composed of fine sands. With sands of the same grain-size, ripple patterns become more curved in stronger winds

(Cooke *et al.*, 1993). In formulating a technique in which ripple profiles and dynamics could be accurately determined, Werner *et al.*, (1986) inferred that the morphology of ripples with respect to wavelength and height was affected by the dune surface slope angle. As elucidated "if the lee-side slipface angle of the ripple is determined by the angle of repose of the sand, one would expect that the slipface angle, relative to the surface of the dune, would become smaller in magnitude for increasing apparent dip" (Werner *et al.*, 1986). However, at four sites over the stoss and crest-brink of a barchan dune, with a constant mean grain-size of 0.27 mm, the ripple windward and leeward slope angles (relative to the dune surface) were contrary to expectation (Werner *et al.*, 1986). Werner *et al.* (1986) suggested that ripple morphology with respect to longitudinal profile, inferred a correlation with dune morphology. Nevertheless, it is most difficult to demonstrate the degree of influence in explaining a ripple's profile as reflecting the dune's surface angle and/or variable shear velocities over the dune-form. It is well documented (Seppälä and Lindé, 1978; Hunter and Richmond, 1988) that given surfaces with constant gradient and sands of constant mean grain-size, a greater degree of sinuosity and longer wavelength results from increased wind shear speed. It is therefore expected that ripple form is a consequence of the interaction between surface angle, shear velocity and direction, and grain-size.

Howard (1977) has shown how the ripple normal to the strike of the crest is not an unbiased indicator of current direction upon sloping sand surfaces. As noted by both Bagnold (1941) and Sharp (1963), the coarsest grains (the creep load) predominantly determine the scale and geometry of ripples. Howard (1977) showed from both mathematical and field evidence, the manner in which the ripple-forming creep load is deflected down-gradient as:

- a function of the surface gradient
- the orientation of the surface relative to the wind
- the friction angle of the sand

Ripple formation on sloping surfaces is therefore due to the coarse grains moving obliquely to the surface wind. For slope angles < 10° the deflection angle is estimated to be c. 1.7 times the slope angle, while for a 20° gradient the deflection angle increases to c. 1.9 times the slope gradient (Howard, 1977). Greatest deflection of the ripple normal from the surface wind direction is particularly high on the horns of a barchan dune (Howard, 1977).

Sorting of sands has been experimentally demonstrated by Seppälä and Lindé (1978) and observed in the field by Bagnold (1941) and Sharp (1963). The finest materials reside in the troughs while the coarsest grain-size is commonly reported to occur upon ripple crests. Seppälä and Lindé (1978) also noted that surface creep fundamentally transported sand to the crest where it remained positioned, until it rolled down the lee slope, assisting in the formation of foreset laminae. According to Sharp (1963) and Tsoar (1990) the process of saltation assists in the creep motion of sand up the windward slope of a ripple, where the rate of creep transport increases upslope from trough to crest. This process produces a ripple surface with a diminution of finer sands, fines being concentrated in the trough between crests, with the ripple body becoming a structure of coarser grade sediment. The lessening of the slope nearing the crest may account for the extreme coarseness of the crest; the diminished rate of creep is not capable of removing coarse grains (Willets and Rice, 1989). Or, as suggested by Bagnold (1941), crestal coarseness is a function of the less energetic saltation impacts for this morphologic position on a ripple. Sharp (1963) however, pointed out that for the Kelso Dunes, coarser grains increase in number upon the ripple, but that the grain-size does not significantly vary on the crest or leeward slopes. In other words, the bulk of the ripple is composed of coarse sand. It is the occurrence of fines predominantly in the trough and part way up the windward slope that signifies grain-size differences, rather than the change in grain-size upslope towards the crest for the ripples of the Kelso Dunes.

Ripple movement is a function of windward erosion and the creation of leeward foreset beds by predominantly coarse materials rolling off the crest. Forward motion vibrates the sands, further distributing finer sands into the core of the ripple (Sharp, 1963), whereby negative skewed, coarse sands dominate the crest, while symmetrical to positively skewed sand resides in the troughs (Binda, 1983). Ripples in uniform sand quickly adjust to changing conditions, whereas those ripples formed in dominantly coarse sands adjust by flattening their crests (Cooke *et al.,* 1993), and may be a consequence of crestal saltation under higher shear velocities (Jensen an d Sorensen, 1986).

2.7 Yardangs and the Process of Erosion

Yardangs are landforms formed by the processes of aeolian erosion - abrasion and deflation. As defined by McCauley *et al.* (1977, 1980), yardangs are sand-abraded ridges of cohesive material that in a classic form resemble the upturned keel of a boat, or as cited by Allen (1982) have a plan-form approximating a lemniscate loop (Fig. 2.8). Ward and Greeley (1984) stated that surface shear would decline as the relative width of the yardang increased and more of the surface was protected. On the other hand, pressure drag would also increase with width, because of the flow separation effects. Where surface shear and pressure drag equilibrate, a planform with a width-to-length ratio of 1:4 develops (Ward and Greeley, 1984; Greeley and Iversen, 1985). However, shape can vary widely, depending on erosion rates, variations in lithology, wind directions and intensities, as well as the maturity of the landform. McCauley *et al.,* (1977) found Peruvian yardangs to have planimetric dimensions that range from 1:3 to 1:10. As with dunes, classification of form is by size, where micro-yardangs are in the order of *c.* 1 m long, meso-forms are between 10-10^2 m and macro-yardangs can reach 10^3 m in length (Cooke *et al.,* 1993).

As commented on by Ward and Greeley (1984) and Laity (1994), comparatively little is known about the genesis and evolution of erosional desert landforms, most studies concentrating on dunes and other depositional features. Yardangs, in particular, have been treated as geomorphic curiosities and have received little quantitative consideration. Although vast fields of yardangs exist in terrestrial deserts, for example those of northern Africa, the Namib of SW Africa, the Taklamakan and central Asian Deserts and Lut desert of Iran (McCauley *et al.,* 1977), relatively recent interest has been stimulated by the discovery of immense fields on Mars (Ward, 1979; Ward and Greeley, 1984). Such presence in both terrestrial and especially in the Martian desert environment, is due to the prevalent uni-directional winds, the intense aridity and favourable lithogies that allow aerodynamic downcutting and sculpturing to predominate. The association of barchan dunes and yardang fields is therefore not uncommon. Nevertheless, yardangs do exist in regions in which winds seasonally reverse, but in such instances, they have one predominant direction and intensity, as found in the Lut Desert (Laity, 1994).

The modification of wind flow through the inter-yardang corridor shapes both the yardang and the corridor, forming either a U-shaped or flat bottomed form (Blackwelder, 1934) which may possess wind ripples (McCauley *et al.,* 1977). The role of abrasion and deflation processes in the formation of yardangs is ambiguous (Laity, 1994), with Ward and Greeley (1984) identifying abrasion as the primary process in cutting the initial landform, with subsequent aerodynamic sculpturing by both abrasive and deflational processes. McCauley *et al.,* (1977) however, demonstrated a contrary view, with deflation as the primary influence of genesis with only minor abrasion acting on the form. The presence of undercutting up to *c.* 1 m height, indicates the apparent limits of abrasion, while the existence of patina atop sandstone yardangs in the Sahara (Laity, 1994) and pre-erosional weathered surfaces for yardangs in Egypt (El-Baz *et al.,* 1979) demonstrates the insignificance of deflation upon the crest of these structures. The identity of slump blocks around the perimeter of yardangs is also indicative of abrasive undercutting (Laity, 1994).

Models of yardang development, unlike those for dune formation, are few. According to Blackwelder (1934), Ward and Greeley (1984) and Laity (1994), yardangs may be initiated in gullies, fissures, faults or joints that are aligned parallel to the direction of the prevailing wind. Abrasion enlarges the fissure, where flow streamlines are compressed, subsequently increasing the flow and shear velocities through venturi action (Whitney, 1983). The bottom and sides of the channel are progressively deepened and widened from erosion that proceeds from the windward corners of the obstacle, to the front slope and crestline, and finally to the leeward corners, flanks and leeward slope, equilibrating ideally with a 1 : 4 width - length ratio (Ward and Greeley, 1984; Greeley and Iversen, 1985).

With such a model, abrasion dominates the windward end of the form, while combined reversed flows and deflation sculpture the mid and leeward regions (Ward and Greeley, 1984). Differential erosion occurs only where the rate of abrasion is relatively low, such as on the mid to leeward portion of the feature (Ward and Greeley, 1984). Velocity profiles of meso-yardangs on the Rogers Lake playa, California, showed Ward and Greeley an accelerating air flow along the windward region, with a decrease in velocity beyond the widest zone of the windward face, and a reacceleration on the leeward ends along the crest and flanks. Erosion of the leeward corners, flanks and the upper surface downwind of the windward slope, is by reverse flow formed by flow separation (Ward and Greeley, 1984). Whitney (1983, 1985) also considered reverse (negative) flow to be consequential in shaping the flanks behind the windward slope for the mega-yardangs of the Kharga Depression in Egypt. Of 27 yardangs measured by Ward and Greeley (1984) at Rogers playa, 24 had a spectrum of width-to-length ratios ranging between 1:1 and 1:10, with sixteen of these twenty-four lying between 1:2 and 1:8 width/length. These results suggested to Ward and Greeley (1984), that yardangs undergo sequential changes of shape in which they become streamlined parallel to the dominant wind direction, developing a planimetry between 1:1 and 1:10. Importantly, the streamlining effect is quite independent of scale in uniform fluid flow, and will alter only when local inhomogeneities such as gullying take effect. Halimov and Fezer (1989), who identified eight morphological classes of yardang in central Asia, promoted minor stages of fluvial erosion as another agent responsible for fissure enlargement on the yardangs in the Qaidam depression.

Ward (1979), using wind tunnel experiments, found that after 200 hours, streamlined landforms developed with a dimensional ratio of length-width-height = 10:2:1. In like manner, Halimov and Fezer (1989) in Asia, found identical dimensions with "low, streamlined, whaleback yardangs". The morphologic variation of yardangs is influenced by:

- rock lithology;
- wind strength, frequency and directionality;
- rainfall and vegetation;

- the amounts of locally available abrasive sediments (Ward and Greeley, 1984).

2.8 Comparative Aeolian Planetology

Terrestrial ergs of crescentic ridge dunes occur mostly on plains of low relief relative to bounding plateau escarpments or mountain ranges. The shapes of these sand seas commonly follows the contours of bordering topographic barriers, e.g. in the Algodones Dunes, California, U.S.A., and the Sahara of Algeria and Libya. Here, numerous basins are filled with dunes predominantly of crescentic and star types, separated by low plateaux and low hills (Breed, 1977). Wilson (1971) has suggested that crescentic dunes are "sand trapping" rather than "sand passing" as with the linear dune, the major axis of which lies parallel to the resultant direction of effective sand transporting winds.

The Martian global distribution of dune sands significantly differs from that of Earth. The ergs of Earth are concentrated in the mid to low latitudes, whereas for Mars, dune accumulations are identified only in the north polar region, numerous crater floors, low-latitude chasma (Fig. 2.8) and near wrinkle ridge settings (Thomas, 1982; Ward and Doyle, 1983: Greeley *et al.,* 1992; Thomas and Gierasch, 1995). Such distributional differences are attributed in part to the global wind patterns, which have resulted in erosion of the equatorial and mid latitudes, and deposition of aeolian debris blankets in the circumpolar regions for Mars (Breed, 1979; McCauley, 1973a; 1973b; Soderblom et al., 1973). The majority of dunes occur between 110°W and 220°W and 77°N and 83°N, while scattered dunes lie between 240°W and 90°W.

2.8.1 Crescentic Dune Morphometry of Hellespontus Region

The first telescopic views of seasonal global obscurations of Martian surface features was interpreted by deVaucouleurs (1954) as massive dust storms. Conclusive evidence of aeolian activity as a geological process was confirmed with the imaging of crescentic dune forms by Mariners 6, 7 and 9 and the Viking Orbiters (Tsoar, *et al.* 1979). Wave-like configurations upon the floors of craters in the Hellespontus region of Mars

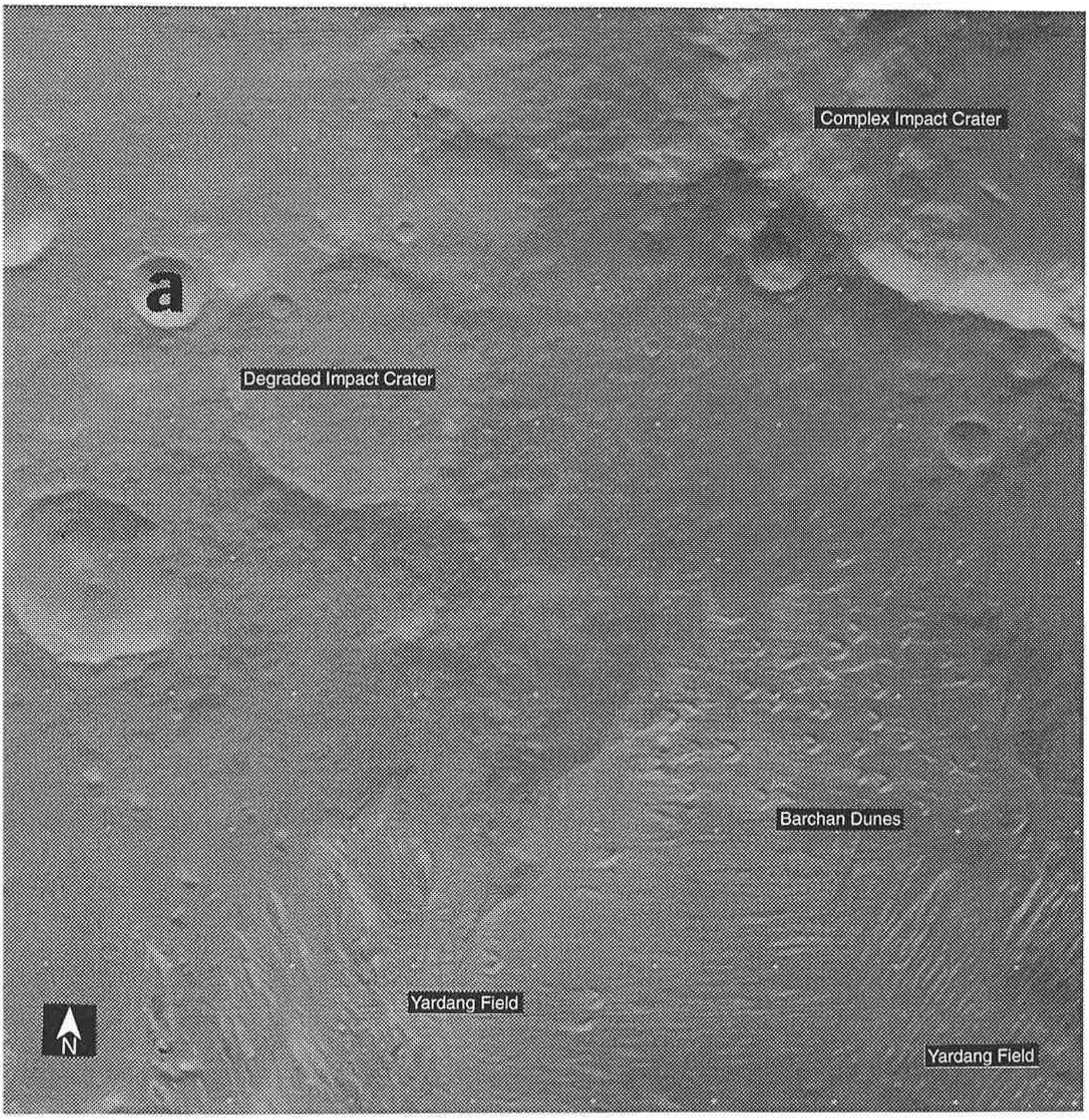


Figure 2.8 A Viking Orbiter I low-latitude, uncorrected, enhanced image of barchan dunes and yardangs within a depositional centre of the Memnonia Fossae region of Mars. Dune migration is westerly, while yardang orientation is NNW-SSE on the western side of the sink and NNE-SSW on the east. This identifies three wind directions, one which orientates the dunes and two which have sculptured the yardang fields. Landform orientation gives insight into both the past and present climatic and meteorological character of the Martian winds. Crater **a** is located at -11° 20', 178° 20' and is approximately 9 km in diameter. (NASA Viking Orbiter Data F438S02 - CD-ROM Volume 8)

(47.5° S, 331° W), were first identified as crescentic dune-forms by Cutts and Smith (1972). The morphology of these ridges appeared strikingly similar to terrestrial barchanoid and transverse dunes, a similarity that was later confirmed by Breed (1977) using dimensional analysis. The scale ratios of width-length, width-wavelength and length-wavelength suggested that the association of size and spacing for the Hellespontus dunes agreed closely with the average relation among terrestrial dunes of the crescentic ridge type. More specifically, comparative examination of orbital images for both Earth and Martian dunefields reveal the Hellespontus region to be similar in geometric form and areal distribution to the Kara Kum Desert in the United Soviet Socialist Republic and Badan-Jiling sand sea of the Gobi desert in China (Breed, 1977).

2.8.2 Crescentic Dune Morphometry of the North Polar Erg

The Viking mission discovered the vast dunefields of the north polar region of Mars (Cutts *et al.*, 1976). The erg comprises an area of 8×10^5 km^2 (Tsoar *et al.*, 1979), while in comparison, the largest active terrestrial erg (Rub Al Khal in Arabia) is 5×10^5 km^2 (Wilson, 1973). Viking Orbiter images identified a concentric swirl of dunes of crescentic ridge type covering an area some 1×10^6 km^2 and circumventing the Martian north polar ice cap between the latitudes of 75°N and 80°N (Breed *et al.*, 1979). Both mega-barchans and smaller individuals are found in the north polar region, with all sizes being similar in morphology to terrestrial barchanoid and barchan dunes. The mega-barchans lag behind the smaller dunes of the field creating a *frame* of large barchan structures (Tsoar *et al.*, 1979). This is typical of the north polar erg and is attributed to the sand saturated winds of the upwind plains depositing their load on the first row of dunes. This first row loses less sand through its horns than it gains from the plains so that the first row turns into mega-barchans. These dunes also grow to a greater height than dunes of the main mass as a result of the additional sand input, and travel at a slower rate than the main field, due to the rate of dune advance being inversely proportional to dune height (Bagnold, 1954; Tsoar *et al.*, 1979). Another significant observation of dune morphology in the north polar erg, is that horn elongation of barchan dunes may show the initial development of seif-linear dunes (Tsoar *et al.,* 1979; Tsoar, 1984).

The terrestrial dunes most analogous to the scale ratio dimensions of length, width and wavelength for the crescentic dunes of the north circumpolar erg (latitude 72° N, longitude 053°) are those of gypsum lithology at White Sands, New Mexico (Breed *et al.,* 1979). Scale analysis shows these dunes to be approximately 1/5 the planimetric dimensions of length and width, than the dunes of the Martian north polar region, although geometric similarity is revealed by the closeness of dimensionless numbers. It is noteworthy that the north circumpolar dunes and those of White Sands reveal less dimensional correspondence to the Hellespontus dunes and their closest terrestrial analogues located within the Peski Karakumy erg. Significantly, these differences express a diversity of geometric form, areal distribution and aeolian dynamics for mid-latitude and polar latitudes on Mars, and the mid-latitude and equatorial erg depressions of Earth (Breed *et al.,* 1979).

2.8.3 Dunes as Indicators of Wind Patterns for the North Polar Erg

Dune classification based on dynamic processes is of two fundamental types, transverse varieties and linear dunes. The advance of transverse dunes is by erosion of the windward (stoss) slope, movement of sand perpendicular to the crestline and deposition on the lee slope (Tsoar *et al.,* 1979), as with terrestrial transverse, barchanoid and barchan structures. Advance of the entire dune body is a result of the primary wind (Tsoar *et al.,* 1979). However dune advance and sediment transport is contrary for linear dunes, where sand transport is parallel to the crestline, and occurs on both lee and stoss flanks, resulting in dune elongation (Tsoar, 1978). However, the idea of transverse dunes originating in only unidirectional winds is incorrect, as winds coming from very narrow field directions (some 10°-20° wide) are rare, except where channelled by topography (Tsoar *et al.,* 1979). Similarly, Smith (1970) has recognised barchans that form in unimodal, bimodal and multi-directional wind regimes. A change in wind direction of 40°-50° is sufficient to begin the elongation of dunes (Tsoar, 1978). Hence, most advancing transverse, barchanoid and barchan dunes in terrestrial deserts have an elongated arm (Tsoar, 1974; Cooke and Warren, 1973).

Approximately one-half of the total area of the north polar erg is covered by transverse dunes with very little underlying bedrock exposed. Modification of the transverse dunes by winds blowing from more than one direction has allowed the determination of wind directions from dune shape and orientation of slipfaces (Tsoar *et al.*, 1979; Greeley and Iversen, 1987). Dunes with elongated horns oriented obliquely to the main direction of advance are typical of secondary winds oblique to the dominant wind direction (Norris, 1966), and coming from the side opposite the elongated horn (Holm, 1960; Tsoar, 1984). Streaks of sediment leaving barchan horns are also characteristic indicators of wind direction. Albedo patterns (crater shadows) and lee-style dune deposits around crater perimeters also indicate principal wind directions. Airflow around craters or hills creates a zone of high wind surface shear (Greeley *et al.*, 1974), resulting in a zone of non-deposition in the lee of a crater/hill.

In a similar fashion, patterns of dunes that fill craters show a V-shaped trough between the dune front and the leeward rim of a crater. This relationship resembles 'echo' dunes on Earth, that develop where dunes meet an escarpment (Cooke and Warren, 1973). From the orientation of these echo dunes, the strongest wind direction can be defined (Tsoar *et al.*, 1979). Besides these regional features, characteristic morphologies of the dunes themselves further refine the gross characteristics of the erg and regional wind patterns. The existence of elongated arms emanating from the linguoid, reversal patterns of crest-lines, and superimposed secondary slipfaces perpendicular to the principle slipfaces, are all indicative (by analogy with terrestrial dune studies) of a variable secondary wind orientation (Tsoar *et al.*, 1979).

2.8.4 Martian Linear Dunes

On Earth, linear dunes are a major dune form accounting for at least half of all dunes in the global sand seas (Breed *et al.*, 1979). It has been suggested that these dunes represent "sand passing belts" for sediment transport across desert basins (Wilson, 1971; Wopfner and Twidale, 1967). By sand motion parallel to the regional resultant wind regime, large volumes of sand are transported downwind to sites of deposition, commonly basin ergs.

On Mars, Viking Orbiter images show an absence of regionally distributed linear dunes, which may be attributed to a lack of abundant sand sources needed to sustain this dune type (Breed *et al.,* 1979). This may imply that much of the available sand has already migrated to sites of accumulation. Without the presence of stabilising vegetation, or a constant volume of sediment input into the aeolian cycle, any linear dunes would eventually disappear by redistribution into the depositional basins (Breed *et al.,* 1979) The absence of linear dunes on Mars is significant. Coupled with the apparent absence of dune sand for much of Mars, it suggests that the Martian sand has been redistributed into a few topographically restricted depositional sites (Breed *et al.,* 1979) by dynamic aeolian processes that are not unlike those active on Earth. Breed and Grow (1979) have suggested that the progression of crescentic dune morphology from simple, compound to complex is a genetic series associated with growth to larger size. This model series has implications for the evolutionary stage of the dunes at each locality, and assumes a similar morphodynamic mechanism for dunes at all localities. If the general assumption of crescentic dune development is progressive from mound, barchan, barchanoid to transverse, then it can be proposed that the Martian localities display mature landforms. Tsoar (1989) adamantly denies that such a relationship is possible for the variants of terrestrial linear dunes, hence caution must also be emphasised concerning crescentic morphologies forming under a single morphodynamic regime.

Contrary to the hypotheses of Breed and colleagues, Tsoar *et al.,* (1979) suggested that the horn-elongated barchans of the north polar erg are seif-linear dunes in an early stage of evolution. These horns mark the initial formation of linear dunes and extend at a more rapid rate than does the advance of the transverse body. With time, these elongated horns dominate the pattern of the dunefield (Tsoar 1978), eventually forming linear-dominated patterns (Tsoar *et al.*, 1979). On Earth, linear dunes are the prevalent type in large terrestrial sand deserts (Jordan, 1964), while transverse dunes command the coastal regions (Cooper, 1958). Terrestrial sand seas such as the Erg Oriental in the Sahara are considered to be $2x10^6$ years old (Wilson. 1971), while coastal dunes are usually < 3000 years old (Goldsmith, 1978). From terrestrial observations of dune types correlated with dune age, as well as from models of dune formation, it is concluded that the dunes of the north polar erg on Mars, represent a relatively young stage of development (Tsoar *et al.*, 1979).

However, if the bedforms of this region are considered immature, it is difficult to correlate this age with the maturity and high degree of organisation displayed by regional megadune patterns. On Earth, such development takes several thousands of years, hence it would be expected that these bedforms would take at least a similar period of evolution in the Martian environment (Breed *et al.*, 1979).

Recently, Edgett and Blumberg (1994), and Lee and Thomas (1995) have reported finding other isolated examples of Martian linear dunes in the region of Noachis amongst two fields of transverse forms and again in the north polar region. As for the linear forms identified by Tsoar *et al.*, (1979), the dunes of Lee and Thomas (1995) appear to be seifs developing from the asymmetric elongation of barchan horns. The mechanism of genesis is suggested to come from very local bi- or multi-modal winds developed by topographic influences and/or channelling. The scarcity of linear dunes on the Martian landscape is due to the unimodal aeolian regime that appears typical for the planet (Lee and Thomas, 1995).

2.8.5 Particle Formation

Viking 1 and 2 and Mariner 9 mission images of the Martian surface showed that aeolian sedimentation is a characteristic feature of the red planet and that sedimentation is associated with landforms of the Martian landscape (Cutts, 1973; Cutts and Smith, 1973;

Sagan, 1973; Binder *et al.*, 1977; Grolier *et al.*, 1976; Sagan *et al.*, 1977; Pollack *et al.*, 1979).

A primary observation of the Hellespontus region was the dark albedo of the Martian crescentic dunes against a lighter background (Cutts and Smith, 1973). The difference in albedo is interpreted to be an important characteristic of the likely compositional differences in dunes on Earth and Mars. Most of the sediments on the land surfaces of Earth are originally derived from the weathering breakdown of granitic rocks to produce quartz sand, silt, clay and the other by-products of feldspars and ferromagnesian minerals (Smalley and Krinsley, 1979). This debris may then undergo further size reduction by particle-particle impact during transport. While terrestrial dunes are mineralogically quartz rich (Smalley, 1966a), Martian dunes are most likely to have a coarse-grained basaltic composition (Hanel *et al.*, 1972; Binder *et al.*, 1977; Carr *et al.*, 1977). This view is further supported by the comparison of similar albedo contrasts as shown by the andesitic barchan dunes of Peru (Hastenrath, 1967; Finkel, 1959). Results of the 'soil' analyses at the Viking lander sites show high concentrations of Fe, S, Ca and Mg (Baird *et al.*, 1976). The relative abundance of these constituents in the regolith point towards a mafic/ultramafic basaltic composition of the source rocks for Martian sediments.

Smalley (1966a) gave a set of constraints operating within the cooling granitic parent, which control the shape and ultimate release of quartz particles. The conditions of high pressure and water content in the magma system from which most granites solidify, suggest that the solid silica forming phase appears as high quartz; the tridymite and cristobolite phases would normally be by-passed in the cooling process. Solid granite develops with high quartz present; this changes eventually by displacive transformation into low-temperature quartz (Smalley and Krinsley, 1979). The high-low transition involves a decrease in volume and results in very high local stresses around each quartz particle. These assist in the breakout of quartz sand particles and contribute to the speed of weathering usually observed in granites. The characteristic size of sand particles is due to events which occur while the granite is still molten. There are no specific external particle forming processes which produce sand size particles (Smalley and Krinsley, 1979). Analyses performed on regolith samples, collected at the Viking landing sites indicated the presence of granitic-acid rocks to be unlikely, and implied that a supply of sand-sized particles originating through a magma fractionating process is improbable.

Nonetheless, Cutts and Smith (1973) have asserted that saltation of sand-sized particulates under Martian meteorological and gravitational conditions does occur. In terrestrial deserts, saltation transports closely graded sands, reflecting the size restrictions predetermined from initial grain formation (Smalley and Krinsley, 1979). Martian sand, however, being deficient in silica would not have such initial constraints placed upon grain size. The sediment must be a weathering product of the dominantly basic lithology of Mars (Baird *et al.,* 1979), and be of a grain-size that can be transported by aeolian processes. Conclusively, it can be stated that terrestrial sands are a by-product of the igneous process whereas Martian detritus is more intimately controlled by the wind regimes and gravitational forces (Smalley and Krinsley, 1979). Experimental work (Greeley and Leach, 1978; Greeley, 1979a; 1979b; Krinsley and Greeley, 1978; Krinsley *et al.,* 1979), has shown that electrostatic forces may bind micro-sized particles together, thus aiding in the control of sediment size at high velocities ($\geq$ 20 m/sec).

Irrefutably, Martian sand must also undergo transportive abrasion and particle-particle impact to produce finer grain, silt and clay, fractions (Huegenin et al., 1978; Fanale et al., 1978; Binder et al. 1977) that would contribute to loess style deposits. Such silt sized aggregates could be similar to the clay-aggregate termed *parna* in the Australian deserts (Butler, 1956, 1974). These clay aggregates can act as silt particles and form loess style deposits. Conclusions derived from the analyses of the Viking landers, point towards iron-rich clays (Baird *et al.,* 1979) being the most likely materials to compose Martian loess deposits.

Impact chipping of larger grains to produce finer fractions is more likely to produce a significant sediment supply on Mars than on Earth, where terrestrial wind velocities are much lower (Smalley and Krinsley, 1979), and glacial grinding by widespread ice sheets has been fundamental in producing silt and clay grade sediment (Smalley, 1966b; Smalley and Vita-Finza, 1968). Martian basaltic materials under wind velocities ≥20 msec^{-1} should produce silt and clay size fractions of quantities not found with impact chipping of quartz on Earth. Observations of the Martian poles indicate that these sites may act as sediment sinks for fine silt and clay sized particles that are incorporated into the polar ice as laminated terrain (Masursky, 1973).

Other sources of sediment for Mars are impact generated debris and volcanic ash (Malin, 1975, 1976; Sharp and Malin, 1975). Krinsley and Leach (1979) have suggested that the origin of fine sediment could be attributed to fluvial transport (Greeley *et al.*, 1977) and subsequent wind erosion. It has been also suggested from laboratory work by Huguenin (1982) that chemical weathering of proposed Martian olivine is a result of H_2O-frost weathering for localities above 40° latitude, where seasonal H_2O-frost accumulates (Wells and Zimbelman, 1989). Besides H_2O-frost being an agent of chemical disintegration, mechanical disaggregation by frost wedging may occur where dust-H_2O condensates occupy pore spaces of fine surficial sediments. Similarly, the high concentrations of sulphate ($MgSO_4$ and Na_2SO_4), carbonate ($CaCO_3$) and chloride ($NaCl$) salts may indicate the physical destruction of rocks by salt weathering processes (Clarke *et al.*, 1976). Here, hydration of the salts or diurnal thermal contraction and expansion of salts trapped in pore cavities, may aid in the disaggregation of the Martian regolith.

2.8.6 Martian Dune Sediments

It has been determined (Sagan and Bagnold, 1975; Greeley, 1978; Tsoar *et al.*, 1979) that Martian saltation and dune deposition require particles in the 0.06mm-2.0mm (~60-2000 micrometre) size range, assuming that the source rock for Martian sediment is basaltic (Carr *et al.*, 1977) in composition and subjected to Martian chemical weathering (Huguenin, 1976). The saltation threshold friction velocity of Mars is an order of magnitude larger than on Earth (see Fig. 2.2 earlier), while the grains most susceptible to saltation have a mean diameter of ~0.16mm (Greeley *et al.*, 1976; Sagan *et al.*, 1977). However, grains of this size in a simulated 5-mbar Martian atmosphere, required surface winds between 90-270km/h for saltation to occur (Greeley *et al.*, 1976). Slope winds of this magnitude are predicted by Gierasch and Sagan (1971).

Further experiments conducted in the MARSWIT (Mars Surface Wind Tunnel) have shown that the most easily entrained sediment grains have a diameter of 0.115mm (Iversen and White, 1982). A threshold friction speed of 25.4 ms^{-1} is sufficient for lifting a 0.1 mm particle of basaltic composition (density 2.5 gcm^{-3}). Surface wind velocities at the Viking 1 landing site reached 17.7 ms^{-1} with gusts of 25 ms^{-1} during the global dust storm of 1977 (Ryan and Henry, 1979). Such velocities are close to the required threshold for saltation at the Martian surface.

2.9 Literature Review Summary

Chapter 2 has reviewed the information assumed most pertinent to this thesis, and has demonstrated that aeolian geomorphology is in its infancy, in both pure and applied settings. A multitude of questions concerning the quantitative geomorphology, sedimentology and modelling of arid landform genesis and evolution exist, and await answers discovered from the integration of further field observations (both terrestrial and extra-terrestrial), wind tunnel and laboratory measurements as well as computer simulations. Fundamentally, aeolian geomorphology (if not all geomorphology) necessitates a multi-disciplinary approach in discovering the facts that will eventually derive truths about this dynamic discipline. To simply describe landscape in qualitative terms, is no longer acceptable.

The following chapters integrate geomorphology, sedimentology, meteorology and statistics in attempting to identify the character, changes and developments of the Gurra

Gurra crescentic dunes. These chapters also consider the processes that are typical of all deserts (both terrestrial and extra-terrestrial), although often as composite mechanisms specific to individual localities.

3 Research Methods

3.1 The Multi-Disciplinary Approach

This chapter outlines the various methods of data collection and analysis used in this study. Various empirical techniques, from several different disciplines, quantitative geomorphology, sedimentology and statistics, formed the basis of this study. These procedures were essential for experimental design, as well as in the collection, analysis and discussion of the morphology and surficial sedimentology of the Gurra Gurra crescentic dunes. Without such a multi-disciplinary approach, the formulation and scientific objectivity of this work could not have been achieved.

The experimental design of this project involved four visits to the field site, one in each season of the year, when both morphological measurements and sediment samples were taken. Field activities spanning all seasons over a quadrennium covered the full range of environmental conditions for observations of the evolution of individual dune members and the dunefield as a whole. Due to logistical constraints, periods of observation ranged between 7 - 14 days. All seasons gave workable data.

Five individual dunes constituted the sample size, and was considered here, to be representative of the crescentic field. Such dune numbers also allowed comprehensive sampling in the time available under all extremes of weather. A survey of the literature showed that most studies incorporated details of only one dune, or at best a less comprehensive comparison of two or three individuals. Where greater numbers of dunes were examined, sampling was less systematic and detailed than that used here, while seasonal comparisons were less frequent, irrespective of the dune numbers. Examples of noteworthy dune investigations are Sharp's (1966) study of the Kelso dunes and Tsoar's (1974;1978) work in the Northern Sinai, with more recent studies by Livingstone (1989a, 1989c) for a linear dune in the Namib, Lancaster's (1989) long term synopsis of the entire

Namibian erg and the work of Burkinshaw *et al.*, (1993) on crescentic dunes in the Alexandria coastal dunefield of South Africa.

Although it was intended that the same dunes should be studied for each visit, the nature of crescentic dunes does not allow all individual dunes to survive independently. Several coalesced with others, or degraded so badly that it became impossible to derive useful information from them. However, in these circumstances, new nearby forms were selected for alternative investigation, with a total of ten different dunes being examined in the four successive years of study.

3.1.1 Dune Morphometry: Data Collection

Slope angles for the lee, crest-brink and stoss elements were measured with an inclinometer placed upon a two metre long base, with an accuracy of ± 0.5°. These measurements were plotted on arithmetic graph paper producing a longitudinal profile of each dune-form for each season. The measurements of length, width and wavelength for the crescentic dunes were obtained from stereoscopic, planimetric views on aerial images at a 1:20,000 scale and a monoscopic field-truthed enlargement of 1:1,670 scale. Measurements were made of forty-two dunes where stoss and lee slopes could be discriminated by spatial pattern and albedo of the interdune floors and dune crests (Appendix 1). To further supplement this two dimensional examination and to allow further comment on the morphology and morphometry of crescentic dunes as previously done by Finkel (1959, 1961), autumnal field measurements of 31 dune lengths and heights were also incorporated (Appendix 2). Height was determined by measuring both linear and angular quantities from the interdune floor to the crest of the slipface on the leeward side of a dune.

Using the formula,

$$y_{opposite} = length_{hypotenuse} \sin \theta$$

dune height was quickly derived, assuming the geometry of a right angle triangle, a shape which equates with the time of sampling in autumn. The calculation of upslope wind-speed amplification used the method of Jackson and Hunt (1975) (see Chapter 2, Section 2.3.2). The aforementioned longitudinal profiles of each dune allowed accurate determinations of dune height, and dune length as measured parallel to the wind at *h/2*, assuming flow was normal to the crestline in each season. This technique facilitated the determination of wind-speed amplification for both normal and reversed winds and, hence wind-speed change over the windward and leeward slopes respectively.

To further understand dune form and process, comparative analysis using previously published data from the regions of the Pampa de la Joya, Peru (Finkel, 1959) and the Salton Sand Sea, California (Long and Sharp, 1964), were also used. Although some mathematical analysis had been previously performed on these data sets, further investigations using like-mathematical and statistical tests (refer to later sections) and like-morphometric parameters have been conducted here.

3.1.2 Dune Granulometry: Data Collection

Although the study of dunes has long been an important part of arid zone geomorphology, the lack of established procedures for dune sediment sampling soon became apparent. In the work reported here, sampling along both longitudinal and transverse profiles enabled a three-dimensional representation of form and surficial sedimentology. It was considered that the granulometric distributions would most probably be influenced by longitudinal transport (either in forward or reverse flow conditions), while secondary transport, although most likely less significant, might occur along the horn slope by either grainfall or oblique air flow. It was expected that the curvature of the dune perimeter would not allow such differences to be easily deciphered from differentiation caused by varying longitudinal distances along the stoss slope.

Because of the intrinsic variability of crescentic dune dimensions during growth, coalescence or degradation, temporal comparisons were constrained to dunes that approached similar size, but which could not have identical dimensions. Due to the assumed controls that dune dimensions may have over the granulometric distributions, statistical analyses derived from small sample numbers have inherent errors. For these reasons, analytical caution was exercised and the use of ordinal data and hence non-parametric statistical methods, became necessary in testing for granulometric change over space and time.

Wishing to discern the dynamic process of sediment sorting upon different morphological locations of the dune, utmost care was taken to sample only the surface laminae. As commented by Sarre and Chancy (1990 p. 357), "samples 'scooped' from the surface have a size distribution that reflects both the surface processes and those of the underlying sediment." Hence, to understand the interaction of surface dynamics and the erosion and deposition of sand, as derived by the variation in sedimentological signatures over the entire dune, it is important to sample only the upper-most layers. This was done by shearing a thin layer of sediment from both ripple trough and ripple crest at random sites over both longitudinal and transverse sections of the dune using a small hand spade. As sedimentary differentiation occurs between ripple troughs and crests, it is important that each sample incorporates a mix of both morphological parts. Samples approximating 50 grams total weight were collected within a 0.5 m^2 surface area, without exceeding at worst, a sampling depth of more than 5 mm in dry sand. More accurate and uniform depths of collection were achieved with the moistened sand of winter. Although not attempted due to the rationing of water for drinking purposes, wetting of the sample points would have enabled more controlled sampling to occur, and it is recommended that future studies use this technique. A collection pattern along each profile gave many sample points within a given morphologic location, over several dunes of similar size. This was deemed to be capable of representing the granulometric distribution for the sand population at that morphologic locality. In this manner, as will be shown subsequently in Chapter 6, the dynamic processes of sedimentary differentiation of the active transport layers was accurately discerned.

To determine the granulometric distributions of the samples, laboratory procedure demanded that the samples be air dried (with any large aggregates, disaggregated by hand), and sieved at one-quarter phi intervals for 15 minutes (see later discussion of these requirements). All moment derivations (see section 3.4) were performed on a PC using the spreadsheet facility of the software package EXCEL, while all statistical derivations (see section 3.6) were performed using the MINITAB package. Because of the mechanical failure of the sieving apparatus (seam separation of the mesh from the sieve aperture) at some undefined time, it was necessary to bin all samples into 0.5 ϕ intervals. Although not all samples were affected by this incident, uniformity of data analysis was considered paramount and was therefore achieved in this manner. Due to the unimodal character of the sediment distributions, whether at 0.25 ϕ or 0.5 ϕ intervals, this manipulation did not detract from the accuracy or detail of the data, but did necessitate the use of non-standard bins when illustrating the distributions.

Due to the non-normal nature of the distributions for some of the morphometric and sedimentological data, as well as, the inability to assess dependent from independent variables in correlation analysis, distribution-free tests were preferred in some instances. These tests enable evaluation of the degree to which two variables vary together and the direction of covariation, but do not go as far as expressing how the variables are related, the form of the relationships, nor their predictive value. Although parametric and non-parametric tests assume quite different data characteristics, the correlation coefficients are not grossly dissimilar for either technique, allowing tentative comparisons between sample sites of other global localities. The bivariate relationships of sedimentological moments were well suited to such analysis. With all morphometric observations, parametric tests were concurrently conducted with those of distribution-free nature, and were shown to identify very little difference between results. Parametric tests were shown to be robust. This allowed standard regression analysis to be used and the resultant relationships tested for significance.

3.2 Dimensional Analysis in Geomorphology

3.2.1 Geometrical Similarity of Landforms

As previously employed by Breed (1977) and Breed and Grow (1979) in the study of dune planimetry, the methods of dimensional analysis and statistical correlation have been applied in the morphometric documentation of the Gurra Gurra crescentic dunes.

Dimensional analysis is a mathematical tool of great importance to quantitative geomorphology, allowing the investigation of geometric, kinematic and dynamic similarities of natural systems and the formulation of rational equations. As in classical Newtonian mechanics, geomorphology extensively utilises the fundamental dimensions of mass, length, and time. All geometrical properties that describe morphology can be reduced to length dimensions, just as all kinematic properties involving velocity or acceleration reduce to the dimension of time, with most being in an inverse relationship with length (Strahler, 1958). In the study of the geometry of planetary surfaces, dimensional analysis is used to assign a dimension to each geometrical attribute of the landform under investigation. Morphological elements such as width, height and wavelength express the dimension of length (L), others such as area have the dimension of the square of length (L^2), while volumetric elements will incorporate the cube of length (L^3).

A different dimensional class includes those morphological parameters which are dimensionless, and are conventionally shown as 0. Mathematically, this means that the fundamental dimensions are raised to the zero power and therefore equal unity ($L^0 = 1$) (Strahler, 1958). An important class of dimensional element is that of angular measurement. For example, the measure of ground slope derived from both vertical and horizontal distances (which incorporate the dimension of length) is expressed by the ratio:

$$\text{tan of slope angle} = \frac{\text{vertical distance}}{\text{horizontal distance}} \underline{\underline{d}} \frac{L}{L} \underline{\underline{d}} 0 \qquad 3.1$$

where: $\underline{\underline{d}}$ means dimensionally equal to

The dimensional parameter *(L)* is dimensionally equal or equivalent to *(L)* of any other similar triangle, because two similar triangles must have corresponding angles equal, whereas the lengths of the sides and the enclosed area can greatly vary (Strahler, 1958).

Another important class of dimensionless morphological parameter is that of combined dimensional parameters, the sum of whose exponents is zero. For example, the length-to-width ratio of a craterform or drainage basin is defined by $\frac{\text{length}}{\text{width}}$ or vice versa. (Strahler, 1958).

Dimensionally this is expressed as:

$$\frac{x}{y} \underline{\underline{d}} \frac{L^{1}}{L^{1}} = L^{1-1} \underline{\underline{d}} 0 \qquad 3.2$$

Or for the element of crater circularity:

$$\frac{\text{measured area of an inscribing circle}}{\text{measured area of a circumscribing circle}} = \frac{\pi y^2}{\pi x^2} \underline{\underline{d}} \frac{L^2}{L^2} = L^{2-2} \underline{\underline{d}} 0 \qquad 3.3$$

Thus any ratio is dimensionless, when it has the same dimensions in the numerator as in the denominator. A dimensionless number can express a morphologic quality that is independent of the linear scale of the landform and consequently has value in making a comparison of morphology for vastly differing dimensions (Strahler, 1958).

"Systems of landforms involving the same geologic processes and materials are generally recognised to possess a considerable degree of similarity. The basis of landform classification depends upon this" (Strahler 1958, p. 280). When conditions of both geometrical similarity (in which form corresponds) and dynamical similarity (in which all forces in the system are proportional) are met, the complete similarity of two systems is achieved. If geometrical similarity exists between two comparative landforms, all linear dimensions are in the same scale ratio of morphologic element and all dimensionless numbers are close to or are identical for both sites. "Combinations of dimensional elements produce numbers that provide descriptive indices of the terrain, irrespective of scale" (Strahler, 1958, p. 279). Such indices evaluate the correspondence of like landforms in an arbitrary scheme of relative magnitude of observed differences, and are termed degrees of correspondence, where the $ratio = \frac{x}{y}$ and x = larger dimension and y = smaller dimension.

3.2.2 Pearson Product-Moment Correlation & Linear Regression Analysis

The Pearson product-moment correlation provides a measure of the relationship between two variables measured on the interval or ratio scale. This was used in both the morphometric and sedimentological studies of the Gurra Gurra dunes. The degree of relationship or association can be expressed in precise terms as a coefficient of correlation r and is given by the formula,

$$r = \frac{\Sigma\, xy}{\sqrt{\Sigma\, x^2\ \Sigma\, y^2}} \qquad 3.4$$

where x is the difference between each value X and the mean of X, $\left(X-\overline{X}\right)$ and y is the difference between each value Y and the mean of Y, $\left(Y-\overline{Y}\right)$ (Cole and King, 1968).

The correlation coefficient always lies between +1 and -1, and when found, allows the expression $100r^2$ to calculate the percentage of variation in y as 'explained' by variations in x, assuming x is the independent variable. Similarly, $100(1-r^2)$ is often used to derive the proportion of variation 'unexplained' or not associated by the regression coefficient.

The values of the two variables when displayed as a scatter diagram, may tend towards a straight line (i.e. the regression line derived by way of the method of least squares). Simple linear regression is a valuable tool of modelling and prediction that algebraically demonstrates the manner in which one variable is related to the other. When the correlation coefficient is zero there is no relationship between the two variables.

Conversely, when the value of the correlation coefficient is significantly high, a relationship between the variables can be suggested. Thus regression analysis offers information regarding the direction of influence (Shaw and Wheeler, 1994). The equations for the regression line,

$$y = a + bx \qquad 3.5$$

where

$$a = \bar{Y} - \frac{\sum xy}{\sum x^2}\bar{X} \qquad 3.6$$

and

$$b = \frac{\sum xy}{\sum x^2} \qquad 3.7$$

The regression equation demonstrates *a* the intercept on the y axis and b the regression coefficient (slope of the regression line as a tangent of the angle of slope). This provides a means of predicting the value of the dependent variable which would be expected for any given value of the independent variable. It is always assumed that the nature of the relationship is a simple linear regression between the two variables, and that equal increments in the predictor variable (x) bring about consistent responses in the dependent parameter (Shaw and Wheeler, 1994).

The estimation or prediction of (y) from the regression equation should be accompanied with confidence limits about the estimate. Confidence limits define the reliability of any estimate of Y, as discerned by the scatter of points or residual variance about the regression line. Also, as the data sample is only a representative of the population, the intercept (a) and regression coefficients (b) for an independent sample group will have intrinsic sampling errors. Hence no one sample group would precisely derive identical estimates of a and b. Each sample will only provide an estimate of the population parameters (Shaw and Wheeler, 1994). Both the inherent variability of the dependent term about the least-squares line and the variability in the regression lines' parameters need to be accounted for. For this reason, confidence limits about the Y estimate are useful and are determined from the equation that denotes the standard error (prediction interval) of any individual estimate by:

$$SE_{\hat{Y}_i} = s_e \sqrt{\left[1 + \frac{1}{n} + \frac{\left(X_k - \overline{X}\right)^2}{\Sigma\left(X_i - \overline{X}\right)^2}\right]} \qquad 3.8$$

where s_e is the standard error of the residuals, n equals the number of observations, X_k the selected X value, $\hat{Y}$ the estimate mean of Y at X_k, and X_i the individual observations of X (Shaw and Wheeler, 1994).

3.2.3 ANOVA Significance Testing for Simple Regression

In general terms, analysis of variance assesses the likelihood of a number of samples having been drawn from the same population. It does so by decomposing the total variance into the variance within each group about their respective group means, and the mean for the whole body of data (Shaw and Wheeler, 1994).

The variance of the dependent variable can be 'decomposed' into that 'explained' by the behaviour of the independent term and into that which remained 'unexplained' (Shaw and Wheeler, 1994). The F-ratio test determines the significance of the ratio of explained to unexplained variance, and provides a result which reveals how effective the regression equation is in accounting for the variability of the dependent term (Shaw and Wheeler, 1994).

The explained variance is termed the regression variance and the unexplained variance the residual variance. For every observed Y value there is a corresponding 'best estimate' value $\hat{Y}$, that is determined by the substitution of the appropriate X term into the regression equation, whilst the corresponding residual is $\left(Y - \hat{Y} = e\right)$ with each Y; $\hat{Y}$ and e have variances. The total variance is derived from the equation,

$$s^2 = \frac{\Sigma(X_i - \overline{X})^2}{N} \qquad 3.9$$

where s^2 is the variance, X_i the ith value of X, $\overline{X}$ the mean X, and N the number of observations.

The regression and residual variances are determined using:

$$s_{\hat{y}}^2 = \frac{\Sigma\left(\hat{Y} - \overline{Y}\right)^2}{k} \qquad 3.10$$

$$s_e^2 = \frac{\Sigma\left(\hat{Y} - Y\right)^2}{n - k - 1} \qquad 3.11$$

where $\hat{Y}$ is the estimated Y values, $\overline{Y}$ the mean of observed Y values, Y the set of individual Y values, n is the number of observations and k the number of predictors ($k = 1$ in simple regression). The F-ratio is then defined as:

$$F = \frac{s_{\hat{y}}^2}{s_e^2} \qquad 3.12$$

The null hypothesis is constructed as one of 'no explanation' for the variability of Y in terms of X. When the two parameters are zero-correlated, the best fit regression line is horizontal ($b = 0$). There is no inter-dependency between the parameters. As b departs from zero, this equality vanishes and some linearity appears in the scatterplots. With a decrease in the scatter about the regression line, the explained variance grows and the unexplained (residual) variance diminishes. The significance of the F-ratio or F-statistic is compared with critical F-values and determines whether H_0 of 'no explanation' is rejected or accepted (Shaw and Wheeler, 1994).

3.2.4 Spearman Rank-Order Correlation

When sample quantities are considered too small to automatically use parametric methods, non-parametric techniques are needed for data description and synthesis. The absence of a Guassian (normal) distribution for a set of observations would necessitate some form of transformation prior to the application of parametric methods of analysis. For example, a positive skew of the data may be eliminated by a $\log_{10}$ transformation into a lognormal distribution. However, the validity of such manipulations are often unreliable with limited (small) sample sizes, and may not necessarily represent the population. With such instances, it is more reliable to use non-parametric (distribution-free) methods. These methods are suitable when the data does not appear to have a normal distribution and can cope with small sample sizes. Such a procedure is valid when dealing with some of the information afforded by the dunes of Gurra Gurra waterhole.

The use of the Spearman rank-order correlation coefficient *(r or r_s)* simply eliminates the need to introduce mathematical transformations and goodness of fit determinations for small data samples. However, using Spearman correlations does not allow confidence intervals for regression lines to be calculated. Spearman's coefficient provides a quantity that is similar and is interpreted the same as the Pearson *(r)* value, with the range between +1 and -1. Unlike the Pearson *(r)* quantity, which is tested for its significance relative to zero using the parametric Student's *t*- test, the Spearman calculation makes no such assumption about the data, and can be accurately applied to non-Guassian samples. Because it is the correlation between the two variables that is of greatest importance, rather than the samples' distributions for this test, the application of the Spearman rank-order coefficient is considered time efficient and accurate for the task at hand. The significance of the correlation coefficient and hence the association of the two variables is done by using Spearman coefficient tables.

The index of correlation is derived from the differences of ranks between the two variables rather than their scores and covariance, as with the Pearson correlation. The rank-order coefficient is found using the formula;

$$\rho = 1 - \frac{6\Sigma d^2}{n^3 - n} \qquad 3.13$$

ρ is the index of correlation between -1 and +1, d is the difference in rank between the variables and n is the number of objects being ranked. Scores that are the same are assigned the arithmetic mean of the ranks they would have received. The correlation coefficients estimated by both the Pearson and Spearman techniques are usually very similar, as shown in Figures 4.3 - 5, Chapter 4.

3.3 Sedimentological Wind Direction Sensors

Surficial sedimentological features formed by air flow over and around dunes have been successfully used in determining antecedent wind conditions by authorities such as Sharp (1963) and Howard (1977). The identification of active seasonal aeolian flow directions for the Gurra Gurra crescentic dunes is also determined by similar methods. Sedimentological features such as ripple forms and ripple migration, and the orientation of sculptured forms including nebkha, micro-yardangs and scour flutes, facilitate the cumulative deduction of air flow direction.

Few investigations have been carried out on the form or characteristic morphology of wind ripples upon desert dunes, although the literature concerning sub-aqueous ripples abounds with nomenclature that assists in defining the degree of sinuosity. Many different morphologies of wind ripples were examined at the site of the Gurra Gurra crescentic dunes, where nearly all forms as defined by sub-aqueous ripple and sub-aqueous dune morphologies (straight transverse, straight swept, sinuous transverse, sinuous transverse out-of-phase, catenary transverse, catenary transverse out-of-phase, catenary swept, linguoid out-of-phase and linguoid in-phase, lunate and interference patterns) were, at some time, observed. Wind ripples have been tentatively graded into three basic types - impact, aerodynamic or 'fluid drag ripples' and granule with each grade highlighting the process of formation (Greeley and Iversen, 1987; Nickling, in Pye, 1994). However, no specific definitions exist for categorising the many shapes that have been observed in each of these categories. The terms straight, sinuous and linguoid for sub-aqueous ripples and straight, sinuous, catenary and lunate for sub-aqueous dunes (Allen, 1982), refer to an increasing complexity of form as flow velocity is increased for a given depth (Leeder, 1982). Likewise, the change of plan-form for wind ripples also has a cognate lineage of shape with change in wind velocity. Therefore, the expression of the degree of sinuosity for the wind ripples of Gurra Gurra waterhole, follows the nomenclature of the sub-aqueous classes, although the processes of formation are obviously different, due to differences in the density contrast (density ratio) between the air:sand interface and water:sand interface.

Ripple patterns for any season are a response to the sand moving winds, their direction, intensity and duration. Variation in gradient upon which the ripple pattern develops, can alter these three parameters and in doing so, also alter ripple sinuosity, wavelength (length) and amplitude (height). Howard (1977) has previously made comment on the bias affiliated with sloping sand surfaces and the orientation of the ripple normal to the ripple crest. By examining a barchan dune within the Salton Sea, USA, Howard (1977) clearly demonstrated that ripple patterns over the horns of crescentic dune-forms are the least representative zones of regional current directions. In contrast, the relatively horizontal inclination of the crest-brink, lower stoss and interdune corridors are favourable localities for use as ripple wind flow sensors. Similarly, Werner *et al.*, (1986) commented on the plausible interaction of ripple morphology and surface slope angle. From these studies of inclined surfaces, it is evident that caution must be exercised in discerning the ripple pattern as indicative of regional air flow direction(s). The dunes of Gurra Gurra show that the progression of one ripple form into another relates to the integration of slope gradient, shear velocity of the wind as well as mean grain-size. Accentuated lines of demarcation between different ripple forms (see Chapter 5, Fig. 5.4) suggest rapid change in the threshold friction speed between different morphologic positions on the dunes. Nonetheless, it is not a simple task to identify whether it is gradient, wind speed or grain-size, or combinations of these, that has brought about morphologic change to the ripple pattern. Although provocative to explore, it is not the intention of this study to resolve the complexities of this issue, but rather to illustrate the many ripple morphologies upon a crescentic dune, and use the most pertinent features formed upon horizontal surfaces, to assist in identifying both regional and seasonal wind directions and activity.

3.4 Granulometric Analysis

3.4.1 The Controversial Phi (ϕ) Scale

Although ambiguity exists concerning the grain-size distribution of dune sands for various desert localities (refer to Section 2.12.1), other ambiguities have been identified with the styles of statistical analysis used in granulometric studies. Techniques involved in the graphic display and numeric analysis of grain-size sorting and distribution, have been shown to possess variations that may be attributed to the technique of analysis, rather than to the natural variation bought about by the aeolian process of dune development. Hence to establish which technique is most applicable to the understanding of dune granulometry at Gurra Gurra, a brief review of the styles of analysis are here presented.

In sedimentology, grain-size is expressed on the Udden-Wentworth grade scale (where particle size in each successive size fraction is expressed as twice the size range of the preceding fraction), and the modified Krumbein (1938) logarithmic phi (ϕ) scale,

$$\phi = -log_2 \frac{d}{d_0} \qquad 3.14$$

where d is the particle diameter and d_0 is the standard particle diameter (1.00 mm) expressed in the dimensionless numbers of phi (ϕ) size. The Krumbein scale illustrates that for each phi-unit increase, the corresponding particle size in millimetres is halved. Because the phi-unit scale is one of dimensionless numbers statistical parameters can be simply elucidated (McManus, 1963; Krumbein, 1964).

Mathematically, the use of a logarithmic scale implies that all particle-size distributions are skewed and unimodal, and that a $-log_2$ transformation transposes the distribution to normality (Gale and Hoare, 1991). Although this is generally not the case, both Middleton (1962) and Blatt *et al.*, (1972) suggest that the approach of grain-size distributions towards normality and the theoretical properties of the normal distribution is a good model to compare observed distributions. Friedman (1961; 1962; 1979) and Folk and Ward (1957) are proponents of the phi scale, they believe that statistical grain-size parameters are environmentally sensitive. Importantly, Krumbein and Pettijohn (1938, p. 241-242) commented that the geometric mean (i.e. metric value of the log value) approximates the centre of a sediment sample more closely than does the arithmetic mean, because it lies "in the cluster of grains near the higher part of the frequency curve. It is thus associated with the most abundant grains in an asymmetrical distribution". To elucidate environment-process comparisons, the transposition of the distribution to a common base $\log_2$ is seen by many workers as reason enough to implement such a procedure.

Nonetheless, Pierce and Graus (1981) in a brief survey of the literature, identified the existence of some misunderstanding about the log function of the phi scale. The non-recognition of the converted logarithmic values being geometric and not arithmetic, is most common. For example, the arithmetic mean of 2.00 mm and 1.0 mm is;

$$1.5\text{mm} = \frac{(2.0\text{mm} + 1.0\text{mm})}{2}$$

where $1.5\text{mm} = \bar{x}$ according to standard statistical convention; whereas, the geometric (antilog of log mean (ϕ)) mean for the units 2 mm (-1.0ϕ) and 1mm (0.0ϕ) is 1.41 mm;

$$-0.5\phi = \frac{(-1.0\phi + 0.0\phi)}{2}$$

where $\bar{x}\phi = -0.5\phi = 1.41\text{mm}$. The geometric value is significantly different when compared to the arithmetic value and should always be identified as such. The suggestion made by Lindholm (1987) and Gale and Hoare (1991) and inferred by Pierce and Graus (1981) and McManus (1982), where sedimentary sizes are given in SI metric units (micrometres, millimetres and metres), may possess some merit in eliminating such errors.

Furthermore, the assumption that the log-normal distribution can accurately perceive process and environmental change, has been questioned. It has been suggested by Bagnold and Barndorff-Nielson (1980) and Barndorff-Neilson and Christiansen (1985) that a log-hyperbolic distribution is more typical of fluid transported sediments (aeolian and fluvial sediments). By using a cumulative frequency curve based upon the log-hyperbolic distribution, Bagnold and Barndorff-Nielson (1980) and Barndorff-Nielson and Christiansen (1985) have been able to differentiate barchan sands deposited by accretion on a dune's crest from those that are deposited by accumulation on the slipface through dune encroachment. Lancaster (1986) and Watson (1986) using the log-normal distribution have also been able to discriminate different sand populations, while Fieller *et al.*, (1984) advocate the use of the log-skew Laplace distribution as does Fleney (1985), who argues that the log-hyperbolic distribution lacks the ability to discriminate sub-populations of sands with several granulometric modal components. Notwithstanding this, Wyroll and Smyth (1985) questioned the superiority of both the log-hyperbolic and log skew Laplace distributions over the log-normal distribution, while McManus (1982) suggested use of the log-normal distribution based on $\log_e$ or $\log_{10}$ (as used by Taira and Scholle, 1977). With 1.0 mm as reference this would represent a larger interval, or a greater ratio of diameters in millimetres, than the phi unit and may well be more suited to resolution of the problem. It is therefore the problem that will influence the method of analysis and determine whether a $\log_2$, $\log_e$ or a $\log_{10}$ transformation is the most appropriate mathematical treatment needed to derive normality or to obtain a foundation from which process differentiation may be observed. It is noteworthy that Pierce and Graus (1981) identified little difference between the geometric means and deviations derived from a $\log_2$ compared to a $\log_{10}$ transformation. However, it may be that these slight variations are significant in process studies and offer greater mathematical propriety in describing grain-size distributions. More recently, Forrest and Clark (1989) suggested that the study of sieved samples, using the multivariate extension to the entropy concept, gives a much "sharper result" for granulometric analysis. Here, samples are grouped in terms of the whole shape of their distributions, thus allowing for variation in the optimal number of interval widths for a particular sample, and permitting the use of more than one set of descriptor variables to be incorporated (Forrest and Clarke, 1989). Nonetheless, Forrest and Clarke (1989) have expressed difficulties with this technique when making comparisons of data from two or more independent studies.

It is evident from these controversies that the perceived universal $\log_2$ normal distribution for aeolian sediments is a mathematical misnomer. However, there is a lack of consensus about any universal method for deriving the distribution most suited for sediments. Krumbein and Pettijohn (1938) referred to both phi and zeta ($\log_{10}$ with a 2.0mm reference) scales as capable units of measurement to simplify and aid in statistical analysis, and clearly did not expect the method of analysis to dictate the problem at hand. The phi scale was devised with the objective of aiding in mathematical simplicity, a function that it still achieves. The use of phi and zeta units were two ways in which Krumbein achieved quantitative statistical analysis of sediments, building upon the descriptive and broad classifications afforded by the grade scales of Wentworth and Atterberg. Moreover, using this pioneering work, Folk and Ward (1957) and Friedman (1962) became dominant figures in applying the phi scale as a significant tool in environmental and process interpretation. However, as Pierce and Graus (1981) stated, many sedimentologists as well as many archaeologists, pedologists, engineers and biologists think in terms of metric units, not phi intervals. Thus, it is paramount that in all research, some comment is made on what phi or any other logarithmic base represents, and that metric statistical parameters, calculated in a logarithmic system, be identified as geometric in mathematical nature. This approach would be definitive in solving some of the problems associated with the phi and log-transformed nomenclature.

Although normality may not be met by a phi transformation, it is still justifiable to use the assumption of log-normality when the numerical indicators of the nature of a distribution are compared with those numerical indicators calculated by other workers over the last half century (Pierce and Graus, 1991). Only when workers begin using and documenting other log bases or log functions (i.e. hyperbolic rather than parabolic-normal), combined with the conversion to metric (geometric) units, will all data be on common ground for comparison, albeit from different mathematical systems. Total abandonment of the phi scale would be a backwards step in sedimentology, as it does offer process differentiation that is not surpassed by any alternate technique so far tested (Ehrlich, 1983; Wyroll and Smyth, 1985). Rather, variations of the log system, combined with co-registration of log values to metric units, need to be fully integrated into modern sedimentological research, starting with recognition of the methodological hazards of particle-size analysis at the undergraduate level of training.

3.4.2 Descriptive Statistics for Grain-size Distributions

The most significant parameters that are descriptive of grain-size distributions and which portray characteristics of the processes of transport and deposition, are the mean, standard deviation and skewness of sedimentary data. Kurtosis is also often calculated, but the significance of this parameter in understanding the geomorphic and/or sedimentologic process is considered negligible (Boggs, 1987). Each of the above parameters can be computed either graphically, using a plot of ordinate probability scale versus cumulative frequency of grain-size, or by the method of moments. The formulae for both graphic (Folk and Ward, 1957) and moment (Friedman, 1962, 1967; Carver, 1971; Friedman and Sanders, 1978; Tsoar, 1978; Boggs, 1987) measures are given below.

As outlined by Swan *et al.,* (1978), error in determining statistical grain-size parameters is inherent with both methods. Firstly, grouping data into size classes and truncating distributions (as with open ended distributions for the silt and clay fractions) introduces inherent mathematical errors. Secondly, the application, particularly of graphic statistical measures (which are defined in terms of a normal distribution Folk, 1974) to non-normal distributions further adds to the mathematical impropriety of the parameters. The computations for the method of moments also assume a normal distribution within each size class (Lewis, 1984), but has some freedom from the bias inherent in assuming a particular physical theory concerning the overall distribution, such as normal, log-normal or log-hyperbolic (Leroy, 1981). Nonetheless, Friedman (1961, 1962) interpreted departure from log-normality as expressed by the skewness of a distribution, as an "environment-sensitive textural parameter" attributed to diagenetic processes affecting the tail ends of the distribution curves. Thirdly, distributions are often bimodal or polymodal in character (Folk, 1971b). Finally, sampling very often involves "bulk" distributions that represent many formational episodes, many of which may be contrary processes of erosion and deposition over time. Grace *et al.* (1978) having examined size frequency distributions within single sand laminae, have demonstrated vertical variation in grain-size. As Gale and Hoare (1991, p. 64) concluded, the log-normal distribution shown by bulk samples is most likely " an artefact of sampling and the mixture of material produced by several formational events." Nevertheless, the application of the phi scale has many advantages and proponents (Section 3.5.1).

3.4.2.1 Graphical Derivation of Descriptive Statistics

The graphical method determines the grain-sizes at which given percentages (f percentiles) of the mass of the sample are coarser. These are obtained from a cumulative frequency plot of cumulative weight percent versus phi grain-size. Although several graphical formulae are available (Trask, 1932; Inman, 1952) the equations designed for a study of the Brazos fluvial environment by Folk and Ward (1957) are those most commonly used.

Graphic Mean

$$Mz_{\phi} = \frac{(\phi_{16} + \phi_{50} + \phi_{84})}{3} \qquad 3.15$$

Inclusive Graphic Standard Deviation (sorting)

$$\sigma_I = \frac{(\phi_{84} + \phi_{16})}{4} + \frac{(\phi_{95} + \phi_5)}{6.6} \qquad 3.16$$

Inclusive Graphic Skewness

$$Sk_I = \frac{(\phi_{16} + \phi_{84} - 2\phi_{50})}{2(\phi_{84} - \phi_{16})} + \frac{(\phi_5 + \phi_{95} - 2\phi_{50})}{2(\phi_{95} - \phi_5)} \qquad 3.17$$

Graphic Kurtosis

$$K_G = \frac{(\phi_{95} - \phi_5)}{2.44(\phi_{75} - \phi_{25})} \qquad 3.18$$

3.4.2.2 Mathematical Moment Derivation

Moment statistics are mathematically superior to the summary statistics of the graphical style, as moments account for the entire grain distribution of the sample. However, the nature of the entire distribution must be known. Any probability distribution has an infinite number of moments, where the first moment (mean) is taken about zero with all subsequent moments being taken about the mean (Friedman, 1962; Middleton, 1962). For a known probability distribution such as the normal or log-normal, the third (skewness) and fourth (kurtosis) moments have definite values, Sk = 0 and K=3. It is therefore possible to compare how close observed and theoretical distributions are to each other. Nevertheless, no individual distribution is defined by a given set of moments, as it is possible to have different probability curves with the same values whether graphic or moment measures are deduced (Friedman, 1962; Middleton, 1962; Blatt *et al.,* 1972).

First moment (mean) $$\bar{x}\phi = \frac{\Sigma f m_\phi}{n} \qquad 3.19$$

Second moment (sorting) $$\sigma_\phi = \sqrt{\frac{\Sigma f(m_\phi - \bar{x}\phi)^2}{100}} \qquad 3.20$$

Third moment (skewness) $$Sk_\phi = \frac{\Sigma f(m_\phi - \bar{x}\phi)^3}{100\sigma_\phi^{\,3}} \qquad 3.21$$

Fourth moment (kurtosis) $$K\phi = \frac{\Sigma f(m_\phi - \bar{x}\phi)^4}{100\sigma_\phi^{\,4}} \qquad 3.22$$

Where, f = frequency percentage of each grain-size grade, m = midpoint of each grain-size grade in ϕ values, and n = 100 percent.

3.4.2.3 A Comparison of Graphical and Moment Methods

The relative merits of graphic and moment analysis of grain-size distributions, and the relationships and compatibility of moment and graphic parameters, have been extensively examined by Inman (1952), Friedman (1962), Middleton, (1962), Folk (1966), Koldjik (1968), Sevon (1968) Jones (1970), Davis and Ehrlich (1970), Isphording (1972), Jaquet and Vernet (1972), Friedman and Sanders (1978), Swan *et al.,* (1978: 1979), Pierce and Graus (1981) and Gale and Hoare (1991) to name only a few who have used one or both techniques in sedimentological work and commented upon the degree of accuracy found. It is not uncommon to find contrary interpretations as to the method deemed most suitable for the task at hand. Many variables are involved in sample collection, the size-frequency distribution and the hypothesis being tested. Accordingly, there exists no one 'perfect' technique of analysis.

Some of the more important considerations are discussed below. The values of summary statistics derived from both graphical and moment methods of grain-size distribution analysis should not be considered comparable, although they essentially derive the same distributional descriptor (Friedman, 1962; Swan *et al.*, 1979; Lewis, 1984; Gale and Hoare, 1991). Davis and Ehrlich (1970), Carver (1970) and Swan *et al.*, (1979) have clearly recognised the often poor correspondence between graphical and moment measures obtained from the same distribution. Variation pertaining to the values of skewness and kurtosis is dramatically evident between the two techniques, while the non-correspondence of all descriptors is greatest for distributions which are significantly divorced from normality (Swan *et al.*, 1978; 1979; Gale and Hoare, 1991).

In determining which style of analysis is 'best' for determining grain-size statistics, both Friedman (1962) and Jaquet and Vernet (1976) advocated neither graphic nor moment measures as being any more significant at accurately deriving representative sample descriptors than each other. Albeit, Friedman (1962) personally prefers the "theoretical elegance" of moment measures and later (Friedman and Sanders, 1978) concluded that moment measures are more sensitive to environmental processes than are graphic measures. Lindholm (1987) also preferred the technique of moments for its superiority to that of graphical methods, in as much as the method of moments can disregard phi (log) notation and directly use arithmetic units. The necessity to derive geometric values is thus avoided. Conversely, however, Lindholm regards the arithmetic standard deviation of little direct significance, "because it cannot be used to compare samples with different mean diameter" For example, a standard deviation of 1 mm doesn't have the same significance in reference to gravel compared to that of sand. Gravel with a mean of 52 mm and a standard deviation of 1 mm is well-sorted, but sand with a mean of 1.5 mm and a standard deviation of

1 mm is poorly-sorted. To avoid this problem, a coefficient of variation must be used (Simpson *et al.,* 1960),

$$coefficient = \frac{s.d.}{\bar{x}}$$

whereas, log-standard deviation allows the comparison of any size. Having examined the attributes and accuracy of both graphical and moments techniques of 100 digitally generated sediment distributions (phi-normal and phi non-normal), Swan *et al.,* (1978; 1979) found that moment measures for both grouped and ungrouped (true) distributions were in close agreement. In contrast, the comparison of graphic and ungrouped moment values for skewness and kurtosis differed in magnitude, but might also infer different environmental interpretations. Nonetheless, the moment method is only of value when the nature of the entire grain-size distribution is known - i.e. it is not open ended (Jones, 1970; Swan *et al.,* 1979). If an excess of fine materials is present for the practical determination of the distribution, then moment measures should not be used (Gale and Hoare, 1991; McManus, 1988; Pettijohn *et al.,* 1987). It is therefore not sufficient to distribute the pan residue (>63 microns) amongst a specified number of classes between the range of 4f ≥ 8f (63 - 4 microns or coarse silt - very fine silt). McManus (1988) suggested that if a total of less than 1% of the sample population is undefined, then errors associated with the use of moment measures are unlikely to be great. However, as the proportion of undefined mass increases, the reliability of the moments sharply decreases (McManus,1988) and graphical measures will be superior in deriving summary statistics about the distribution.

It was also demonstrated by Swan *et al.,* (1979) that truncated samples where all fine pan residue was 'lumped' within a given interval, provided far more reliable results than did graphic approximations for skewness and kurtosis. Moreover, all moments become less accurate as more material is included in the unanalysed pan fraction. It was found that the least accurate results for skewness and kurtosis were derived when pan fractions approached between 5%-10% of the total weight. Earlier work by Jones (1970) using the Edgeworth series of moment measures, found that a pan fraction of 1% of the total sample introduced error, whereas up to a 5% truncation for the graphic method is acceptable.

Another significant consideration in regard to grain-size analysis is the phi interval used to derive the sediment size fractions. Pettijohn *et al.,* (1987) and McManus (1988) commented that the one-quarter phi interval is recommended in grain-size analysis, while Moiola and Weiser (1968) showed that graphical statistical measures derived from grain-size data sieved at one-quarter phi intervals were more effective in discriminating depositional environments than for measures derived from the same data at whole or one-half phi intervals. Contrary to this, Swan *et al.,* (1979) suggested that whole grade intervals are as accurate at determining graphic measures as are finer intervals. However, one-quarter phi intervals are shown to be an improvement in accuracy for statistically derived moments and decrease the inherent error of grouping sediment data in size fractions, a feature that Folk (1966) associated as a significant error factor, associated with the method of moments. Similarly, Hails *et al.,* (1973) have demonstrated that moment measures approach their true value more closely when derived from one-quarter phi intervals than either one or one-half phi spacings. It was noticed from this work that as the phi interval increases, the second moment (sorting) also increases, which indicates a slightly poorer sorting characteristic. However, it is the third (skewness) and fourth (kurtosis) moments that are most sensitive to small changes in the grain-size distribution data (Hails *et al.,* 1973). Jones (1970) and later Swan *et al.,* (1979) explained the greater sensitivity of the higher moments as being related to the tails of grain-size distributions. The largest contributors towards skewness and kurtosis are the coarse and fine fractions, these being of low weight per cent and most distant from the mean. It is for these reasons that graphic measures are deemed less accurate with non-normal distributions, especially for skewness and kurtosis, as compared with moment measures, as the latter account for the entire distribution (especially at one-quarter phi intervals), while graphic methods only take into account a range between the 95th and 5th percentiles.

With due consideration, the method of moments was chosen as the most reliable technique from which the sediment distributions of the Gurra Gurra crescentic dunes could be determined and accurately indicate the style of sorting process.

3.5 Granulometry and the Geomorphic Process

Statistical measures are valuable as indicators of the environments (processes) of transport and deposition of sedimentary particles (Folk and Ward, 1957; Friedman, 1961, 1962). The mean (a measure of central tendency) of the grain-size distribution reflects the average kinetic energy of the depositing medium, while the standard deviation (sorting or dispersion), is a parameter that is dependent upon velocity variations and bimodality of the sediment source (Pettijohn *et al.*, 1987). The measure of a grain-size distribution's asymmetry is suggestive of environmental settings (Pettijohn *et al.*, 1987) where skewness ranges between +1.0 (positive or fine-skewed) to 0.0 (symmetrical) to -1.0 (negative or coarse-skewed). Asymmetry, as expressed by the tails of the distribution, is considered to be environmentally sensitive (Pettijohn *et al.*, 1987) and can therefore infer aspects about the transport processes (environmental significance) involved with the grain-size distribution (Pettijohn *et al.*, 1987). Kurtosis measures peakedness of the grain-size distribution, or more specifically, the ratio between the sorting of the 'tails of the grain-size distribution and the sorting of the central portion of the distribution (Lindholm, 1987). Flat peaked curves are termed platykurtic while sharp peaked distributions are said to be leptokurtic. Although calculated with the first, second and third measures, the geological significance of kurtosis is not well understood (Boggs, 1987) and has been little used to derive geologic conclusions (Pettijohn *et al.*, 1987). According to Boggs, (1987) it is assumed to have little value in interpretative grain-size studies.

Tsoar (1978) commented that grain-size characteristics of dune sand show valuable information about dynamic processes and that grain-size assists in determining dune morphology, i.e. whether high or relatively flat forms develop. By sampling the sand layer subjected to aeolian activity (ripple laminae) within distinct morphological zones across the dune (lee, crest, brink, stoss), it is possible to study the differences in the grain-size characteristics and to identify their influences on dune morphometry (Tsoar, 1978) and the processes of erosion, entrainment and deposition of dune sands. It is this style of analysis that will constitute the comparative sedimentological and morpho-dynamic study of crescentic dunes in the Strzelecki Desert, South Australia.

3.6 Experimental Design

Both the literature review at the commencement of this study, and later direct field observations, identified that upper dune dynamics are grossly different form those which occur on the lower parts. Consequently, partitioning of the gross morphologic element was demanded. To define and enhance differences in the spatial and/or temporal variation of sedimentary moment measures for the morphologic elements (lee, crest, stoss) of the dune, locations were partitioned for each, into higher and lower regions (i.e. low lee, crest-brink, low stoss, low horn east and low horn west). By using this approach, all seasons and morphologic locations had data samples where $n \geq 5$, and thereby permitted the use of distribution-free methods of data analysis (refer to Section 3.6). Hence, the crestal region contains samples from the upper half of the dune elements (stoss, lee, horn east, horn west), while the lower micro-elements are from the lower half of each individual dune element. Further partitioning and hence statistical testing of three units within each element was considered, however, due to insufficient sample numbers this was considered improbable.

3.6.1 Hypothesis Testing of Grain-size Distributions

As outlined by Siegal (1956), there are many advantages in using non-parametric statistics, especially when the data is of a small sample size and the distributions cannot be assumed, or be tested for, normality. This approach does not diminish the accuracy nor reliability of the samples, but of course, it is statistically desirable to have large sample sizes as the likelihood of making Type I or II errors decreases as sample size increases.

It was found that pooling of the sedimentological data, for at least three dunes gives an acceptable indication of both spatial and temporal change. Nonetheless, dune numbers of five or greater are intuitively considered more appropriate sample sizes representing the Gurra Gurra dunefield. Just as important, are the numbers of samples taken for each dune. Two non-parametric tests;

- the Kruskal-Wallis test for three or more independent k-samples
- the Scheffé-type projection for differences between three of more independent k-samples

were used to test for 'difference' between the bivariate relations of dune morphology and sediment distribution, and sediment distribution and season. Furthermore, the testing of significance for the test-statistic(s), made no assumption regarding directionality for any of the granulometric variables. The alternate hypothesis would be erroneous to subscribe to a given direction of grain-size, sorting or skewness, when complex and possibly contrary interactions between morphologic elements, sedimentology and seasons occur. Hence the two-tail test of significance is used to accept or reject the null hypothesis of 'no difference'. The use of a two-tail level of significance enables a meaningful deliberation of whether mean grain-size, sorting and skewness is different between the morphologic and seasonal variables.

3.6.2 The Kruskal-Wallis H statistic

The Kruskal-Wallis test is a one-way analysis of variance using ranks. Individual grain-size measures are transformed to ordinal data (ranked) and classified into groups of major morphologic elements (lee, crest, stoss, east and west horns). The significance of inter-group contrasts are established using sample groups of size [k=5], and are shown by the size of the H-statistic. This test assists in determining whether grain-size, sorting and skewness for each morphologic sample, could have derived from a common population. The Kruskal-Wallis test is most effective as a preliminary analysis of the difference between the different depositional environments (morphologic elements). Nevertheless, an H-statistic of significance gives no precise information about the nature of differences between the environments. Rather, H solely tells whether a significant difference between morphologic elements is found and whether a more detailed analysis of differences between pairs of morphologic elements is warranted. The H-statistic is calculated on the sum of the ranks for each sample when all observations are placed in rank order (Matthews, 1981). If all k-samples have been drawn from the same population, then the sum of the ranks for each sample would be approximately the same. H is therefore a measure of the actual difference in the sum of the rankings between samples. If this difference exceeds the difference that is likely to have occurred by chance, then the hypothesis of 'no difference' between k-samples is rejected (Matthews, 1981).

The H statistic is given by the formula:

$$H = \frac{12}{N(N+1)}\left[\sum_{i=1}^{k} \frac{\overline{Ri}^2}{n_i}\right] - 3(N+1) \qquad 3.23$$

where N is the total number of observations, $\overline{Ri}$ is a set of rank sums for each k-group and n_i the number with each group.

Where there are greater than five observations in each group, the test statistic is distributed as chi-square (χ^2) with k-1 degrees of freedom. Also noteworthy, is that where inter-group differences are large, the H statistic is also large, and is useful when wanting to find k-groups that have greater difference from each other.

With mean grain-size data, a high proportion (> 25%) of tied raw values were encountered. To aid in eliminating distortions caused by ties, a correction is made to the H statistic where:

$$C = 1 - \frac{(T^3 - T)}{(N^3 - N)} \qquad 3.23$$

T is the number of ties and N is the number of observations. The corrected H statistic = HC.

3.6.3 The Scheffé-type projection

The Kruskal-Wallis test rejects H_0 if H exceeds the critical value. In such cases, a difference is expressed between k-groups, however, no indication is shown of which k-groups are different. To identify which data groups (morphologic elements) are different, the application of a Scheffé-type projection (Kolz & Johnston, 1985) extends the quantification expressed by the Kruskal-Wallis test, where:

$$\left|\overline{R}i - \overline{R}j\right| > \left(\chi_{k-1}^{2\alpha}\right)^{1/2}\left[\frac{N(N+1)}{12}\right]^{1/2}\left(\frac{1}{n_i}+\frac{1}{n_j}\right)^{1/2} \qquad 3.24$$

k= number of groups, N = total sample size, n_i, n_j = sample size for the groups and $\left|\overline{R}i - \overline{R}j\right|$ is the test statistic.

As with the Kruskal-Wallis test, the Scheffé-type projection requires that $n \geq 5$ in each k-group. Where the test statistic $\left|\overline{R}i - \overline{R}j\right|$ is greater than the critical value, it can be assumed that a H_0 of 'no difference' is rejected, and a significant difference that exceeds the difference that is likely to have occurred by chance, exists between k-groups.

3.7 Summary of Research Methods

This chapter has outlined the various empirical techniques used in this study. Both qualitative and quantitative observations have been made on the morphology and sedimentology of the Gurra Gurra crescentic dunes. The proceeding chapters demonstrate the need to combine both methods to deduce the morpho-sedimentary identity and processes active in the evolution of these dunes. As will be demonstrated, in the determination of dune sedimentology, graphical illustrations (see Figs. 6.1-6.31, Appendix 3) show general trends between dune morphology and sediment distributions. However, these general trends offer no concrete information about the relationships between variables. One must question whether observed trends are strong enough to be considered real. It is for this reason that statistical analysis is indispensable in determining whether the observed patterns are 'real' or coincidental.

Nevertheless, it is prudent not to expect statistical results to be an end in themselves. Without rigorous deductive reasoning behind the interpretation and validity of test results, statistics can be worthless. On the other hand, the purely visual interpretation of graphed data can also lead to incorrect conclusions from which unreliable models may be derived. The use of both graphical (qualitative illustration) and statistical tests allows stronger conclusions to be drawn. If reliable models of dune development and mechanisms of aeolian process-response are to be developed, the application of quantitative methods to seasonal sampling of many dunes and their micro-geomorphic environments is essential for understanding the operation of aeolian systems. Even with careful fore-planning, hindsight has demonstrated some methods to be not as accurate nor as useful as others. It is also apparent that very few, benchmarks exist in the analysis of aeolian landforms, and that those that do, may not be as universally acceptable as often presumed. For example, the ubiquitous usage of the lognormal (phi) scale in analysis of aeolian sediments has to be reassessed. Nevertheless, the methods of research used here, have detailed the manner in which spatial and temporal variation of the geomorphology and surficial sedimentology of the Gurra Gurra crescentic dunes has occurred.

4 Crescentic Dune Morphology and Morphometry

4.1 Dune Morphology and Regional Wind Patterns

This chapter examines seasonal variation for the morphological and morphometric character of the Gurra Gurra crescentic dunes. The variable morphologic character of the crescentic dune-field is a response to seasonal wind patterns that are locally enhanced by the shape, size and inclination of the slopes of the dune-forms. Field observations of both regional and intra-dunefield[1] wind directions are accurately reflected by the direction of dune advance, slipface orientation, the position of lee projections and depositional aprons, the streamlining and fluting of meso-yardangs and nebkha, obstacle-induced erosional moats, as well as the orientation and overprinting of ripple pattern forms. Combined, these observations provide insight into the directional variability of the seasonal winds at both the regional and local intra-dune scale. In like manner, the variability of the seasonal winds correlates with the morphological variability of the crescentic form. This association generates an understanding of the morpho-dynamic mechanisms involved in the genesis and evolution of the crescentic dune shape.

The changes observed in morphology of the Gurra Gurra crescentic dunes correlate with annual fluctuations in the regional wind regime as displayed with the sand roses of Fryberger (1979) for the Simpson Desert (see section 1.3). A broadly trimodal potential for sand drift in the Simpson Desert is significant in that dominant summer (Dec. - Feb.) south-easterly and winter (Jun. - Aug.) northerly and south-westerly wind vectors respectively coincide with the observed coincident and reversed morphology of the dunes respectively. During August through to September (spring) the wind trend shifts predominantly to the south-west and roughly coincides with the vehement winds that blow across the Gurra Gurra crescentic dunes from the WSW to the ENE, causing oblique saltation in the afternoon hours and the development of a second stoss slipface. Fryberger (1979) dismissed the northerly mode of effective sand-moving winds as insignificant because they

[1] The intra-dune is the local corridor between crescentic dunes developed upon the underlying linear dune, as opposed to the interdune (swale) between linear dunes.

comprise only 12% of the total drift potential and occur during the time of weakest wind intensity. However, field observations at Gurra Gurra waterhole reveal that the winter mode of northerly sand drift potential is the weakest and least effective saltation period of all seasons. Saltation is substantial when winter storm fronts cross these regions. Furthermore, the effect of shear velocity amplification upslope of the flanks of both the linear and crescentic dunes accentuates this influence. Northern storm front activity coupled with the modest but persistent northerly winds are responsible for developing the morphological and sedimentological distributions observed during the winter period. As stated by Hastenrath (1967 p. 329), and demonstrated by the crescentic forms at Gurra Gurra waterhole, "barchans act as extremely sensitive anemometers".

4.2 Dune Morphology

Planimetry of the dunes surveyed from aerial photographs, demonstrates a mean length of 25 m, a mean width of 22 m and an average 40 m wavelength between opposing crests with a mean height of 3.5 m. However, observations of the study dunes suggest that the dimensions are seasonally variable (Fig. 4.1). For example, the average size of the autumn dunes are 31 m long, 24 m wide and 3.4 m high, while those of spring have an estimated mean length of some 40 m. Longitudinal profiles (Fig. 4.1) show that this cyclical change in morphology and morphometry is a response to seasonal changes in the wind regime. By maintaining constant height, these diagrams illustrate the changes in dune form and length as characterised by the seasonal wind regime.

Figure (4.1a) illustrates the lee to stoss, convex-to-concavo-linear, winter longitudinal form developed under the influence of dominant northerly winds. A leeward apron inclined at between 2° - 7° extends some 2 m from a gentle convex slipface. The slipface ascends from a maximum 28° at the lower lee, gradually changing to < 10° at the upper lee-brink position. The crestal zone is sub-horizontal with the lee at a median inclination of 17°.

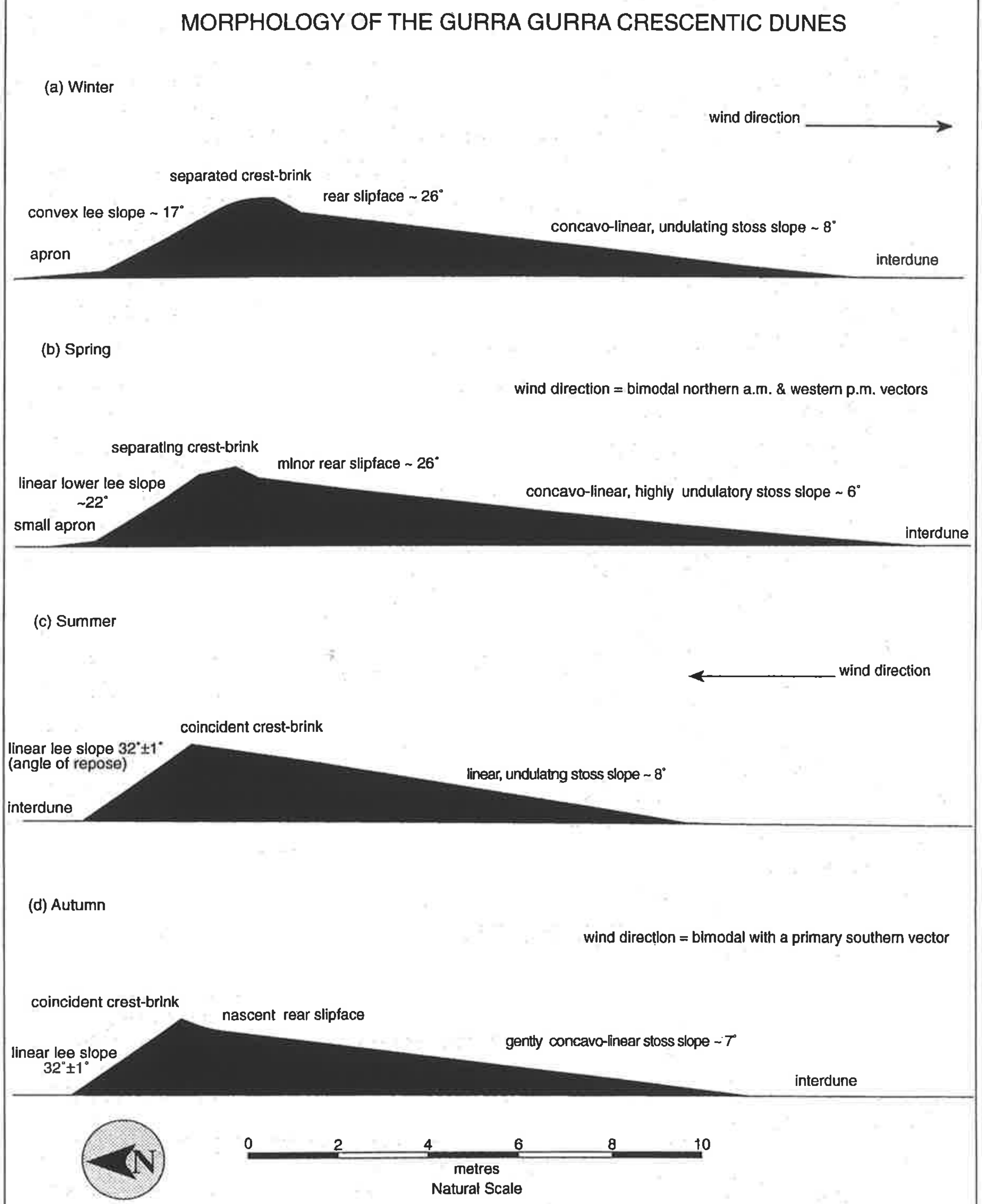


Figure 4.1 Schematic longitudinal profiles of the dominant morphologies of the Gurra Gurra crescentic dunes. The winter shape demonstrates a reversing form originating in a northerly wind, while the summer morphology develops under the influence of southerly winds, and characterises the equilibrated form for crescentic dunes under the influence of uni-directional winds and intensities capable of dune migration. The morphology of spring is intermediate between those for winter and summer, while the autumnal shape in mid-April was prevalently summer-like but for the presence of a nascent rear slipface on some dunes identifying the presence of minor reversing winds. Due to migratory change in the crestal position, minor variation in height occurs, but not as noticeably as the change in dune length. Maintaining a constant height, the length to height ratios are shown as ~10:1 in winter, slightly shorter in autumn ~9:1 and summer ~8:1, and slightly longer in spring ~11:1. All angular quantities are expressed as the median of each micro-morphologic element. However, the stoss slope does not exceed 15˚ nor does the lee slope exceed 34˚.

A separated crest-brink forms a relatively wide zone with minor dune lowering. The crest acutely transcends windward into a rear slipface inclined at 26°. The stoss slipface is consistently ≤ 1 m high and develops 'hour-glass' patterns with sand avalanches and 'crestline-notch' effects. The windward gradient has a median slope of 8° gently rising from the sub-horizontal intradune floor and does not exceed a 15° inclination. Undulations or variation in gradient over the stoss slope are consistent with the position of ripple sets of varying dimensions, illustrating the integration of slope inclination, grain-size and ripple pattern. The ripples reveal approximate flow direction and wind intensity upon the stoss element. They further demonstrate the characteristic reversed wind pattern of winter, with the lee of the ripple facing southward (also refer to later section 5.0.1). The mean height-to-length ratio of 1:10 characterises the non-equilibrated crescentic form of dune reversal and windward extension.

Summer form is in direct contrast to the morphology of the winter dunes (Fig. 4.1c). High intensity, long duration, southerly winds sculpture and advance the summer dunes, when both saltation and significant suspension are active. The leeward sand apron is absent, while the slipface is at the angle of repose (32°±1°) from crest to intradune. A sharp intersection of the lee and stoss slopes forms a coincident crest-brink line, which defines a mean height:length ratio of 1:8. This is significantly different from that of the winter dunes and shows that dune height is only one-eighth of dune length, a morphology that corresponds with the forward stance of the crestline and northward advance. The stoss is linear with a gradient not exceeding 14°, with zones of ripple-induced undulation and gradient variation that characterise a median slope grade of 8° (Table 4.1). Ripple patterns and migrations identify a prevalent northerly wind orientation (see later section 5.0.2), while the absence of the rear (stoss) slipface is also characteristic of the summer dune form, demonstrating morphological and hence dynamic equilibrium (also see later, Fig. 5.1).

The autumnal dunes reflect the morphologic character of the summer dunes. During mid-April the winds are light and rarely reach saltation velocities. The slipface angle is at a maximum (32°±1°), with the maintenance of a crest-brink coincidence. However, some dunes display development of a nascent stoss slipface inclined at 26°. The ratio of height:length does not significantly alter in early autumn from that of summer (H:L ~ 1:9). However, the minor changes in the upper micro-morphologic configuration of the dunes is indicative of the cyclic response of a change in the seasonal wind regime as winter conditions approach with a prevalent reversed flow regime.

Dune morphology in late spring displays characteristics typical of both summer and winter conditions (Fig. 4.1b). Considerable morphologic differences are expressed that reveal spring to be an aeolian regime intermediate between quasi-equilibrated morphologies of the dominant southerly wind directions of summer, and the northerlies of winter. Not unlike winter, an apron of lee side deposition prevails on those dunes whose gradients are not at the angle of respose. This reflects either creep of the basal sands on the lower lee flank and/or avalanche of sands from the migrating crest.

Morpho-Element	n*	Mean°	Median°	Minimum°	Maximum°
Winter					
Lee	35	14.5	17.0	0.0	29.0
Stoss	49	8.0	8.0	1.0	15.0
Horn-East	33	5.0	4.0	0.0	15.0
Horn-West	11	8.5	8.0	1.0	20.0
Rear Slipface	06	26.5	26.0	25.0	28.0
Summer					
Lee	16	32.0	32.0	30.0	34.0
Stoss	55	8.5	8.0	4.0	14.0
Horn-East	65	6.5	7.0	0.0	16.0
Horn-West	38	7.0	6.0	0.0	28.0
Autumn					
Lee	29	31.5	32.0	30.0	34.0
Spring					
Lee	33	19.0	22.0	0.0	33.0
Stoss	82	5.5	6.0	0.0	13.0
Horn-East	36	6.5	7.0	0.0	16.0
Horn-West	52	10.5	10.0	2.0	23.0
Rear Slipface	10	25.0	25.5	18.0	32.0

Table 4.1 Descriptive statistics of slope gradients for the Strzelecki crescentic dunes. Note n* is the number of measurements over the dune element and not the number of dunes sampled.

Although the leeward slope for some dunes is 32°±1°, many more gradients are between 24° - 28° (median = 22°) (Table 4.1), with the upper lee slope becoming convex in profile as the crest migrates northward under reversing winds. Overall crest-brink separation is prevalent, but not as differentiated as with the winter dunes. Rather than a broad flat crest, the crestline rises to an apex which mimics the antecedent summer shape (Fig. 4.1b). Mean dune height-to-length ratios portray a slight decrease in height relative to length, with H:L ~1:11.4. The development of the rear slipface inclined at 26° with a height of ≤ 1 m, is a feature that mirrors the form of early winter. Just as significant, however, is the presence of a less well developed but prominent second rear slipface, orthogonal to the first on the eastern flanks of the dunes, and dipping as with all other stoss slip slopes, at an angle of ~26° (Table 4.1). This feature infers a response to an oblique westerly air flow and demonstrates that reversed slipfaces at these dimensions reach a slope maximum of approximately 26°. Ripple patterns (see later section 5.0.4) are indicative of diurnal bimodal wind directions. The inclination of the stoss gradient is characteristic of all other seasons with a low median value of 6°.

The morphologies of the Gurra Gurra dunes are responses to seasonal variations in the wind regime. Crescentic morphology is cyclical and continuous between stages of very short-lived equilibrium and different phases of quasi-equilibrium. The time of observation is therefore important in the perception of dune shape between the stages of equilibria. Figure 4.1a-d is a 'snapshot' of dune shape at four discrete reference points along a continuum of almost infinite but mostly subtle change. Such illustrations approximate the seasonal form of the Gurra Gurra dunes. Nonetheless, recognition of shape difference can be postulated by subdivision of each major temporal unit. For example, the winter, spring and summer profiles are illustrative of mid-season, when the aeolian regimes of opposing seasons are least influential. However, Fig. 4.1d is a profile indicative of early autumn and transient influences of the previous summer. In other words, the summer shape was not erased at the time of observation. However, as autumn progresses, winds re-sculpture the equilibrated summer form into a quasi-equilibrated, retrogressed morphology that merges with the reversed quasi-equilibrated form of winter. Progressively, the reversed form enters into another stage of quasi-equilibration under the spring regime, with the transgression of crestal features and the shortening of dune length towards the equilibrated crescentic shape of summer. It is during this time frame that both barchan and compound dunes demonstrate classical crescentic form and migration.

Significantly, dune height varies little, while dune length shows greater seasonal contrast. The Gurra Gurra dunes apparently undergo a mass transfer whereby the windward slope expands and contracts, but alters the crestal height of the dune very little. The reversed form of winter, and the transitory spring form, have the greatest windward elongation compared to the lesser lengths of the autumn and summer forms (Fig. 4.1a-d).

The changing morphologies of the Gurra Gurra dunes reflect simple dune-form relationships. The summer stoss-to-lee, linear-linear profile increases in height, while dune length decreases. Mass transfer occurs vertically along the *z* dimension with a reduction in the *x* plane. Although not quantitatively determined, dune mass is not significantly lost, but redistributed under the influence of prevailing southerly winds (Fig. 4.1). Contrary to this, the winter shape demonstrates that mass is redistributed longitudinally with a slight reduction in dune height. This form is quasi-equilibrated with the direction and intensity of reversed northerly winds. If these winds were of greater intensity and/or duration, a total reversal of form would be instigated eventuating in an equilibrated coincident crest-brink and shortened shape. Spring is indicative of the greatest height reduction and greatest increase in dune length, as two intense diurnal wind directions assist in elongating the stoss slope. Such elongation and mass transfer may be assisted by lee slope amplification in both spring and winter, with greater amplification and transport occurring on the slightly longer length and steeper grade of the vernal lee slope. It is apparent, however, that greatest change to form occurs upon the crestal region of the dunes.

This conceptual model encapsulates the interrelationships between reversing and oblique wind directions, wind intensity and duration, and dune morphology and morphometry.

4.2.1 Wind Obliquity and Dune Morphology

Two principal orientations of gross form are reflected in the Gurra Gurra crescentic dunes for all seasons and are illustrated in the morphographic map (Fig. 4.2) of the southern half of the crescentic dunefield. Although seasonal morphologic change transpires, especially to the upper dune, the meso-dimensions of the Gurra Gurra dunes are relatively stable and are able to withstand the minor seasonal variability of the aeolian regime. A distinct trimodal annual wind regime of WSW, SSE and NNW is maintained with these dunes, with the SSE being the primary component, the WSW the secondary component and NNW the tertiary vector of influence. The gross elongated crescentic shape of the dunes is derived from the primary and secondary directions, with the third component only influencing the upper morphology.

Elongation of the eastern horn of individual barchans as well as the elongation of the linguoid element for compound ridges, forms seif linear dunes and oblique transverse ridges, respectively. The mean orientation of the longitudinal axis and the direction of advance for the dunes is 330° (*c.* NNW) (n = 44). The crescentic shape of the dunes is maintained and oriented towards the NNW, with an oblique skewness of gross form deliberated by the elongation of the eastern horn *c.* 20° from the longitudinal axis. Figure 4.2 illustrates the process of horn elongation via the action of a bi-directional air flow. Elongation into a seif linear is restricted to individual dunes (barchans) where the intradune area is greatest between landforms and where obstruction to easterly air flow is at a minimum. Such dunes often involve the coalescence of individuals into the elongated path of slower moving, larger dunes, the smaller masses becoming intercalated with the development and growth of the horn into a seif linear dune. Others, however, show only the elongation of the horn, which will inevitably interact with and more often than not be absorbed into barchanoid or transverse dune ridges in this closely spaced dunefield (see

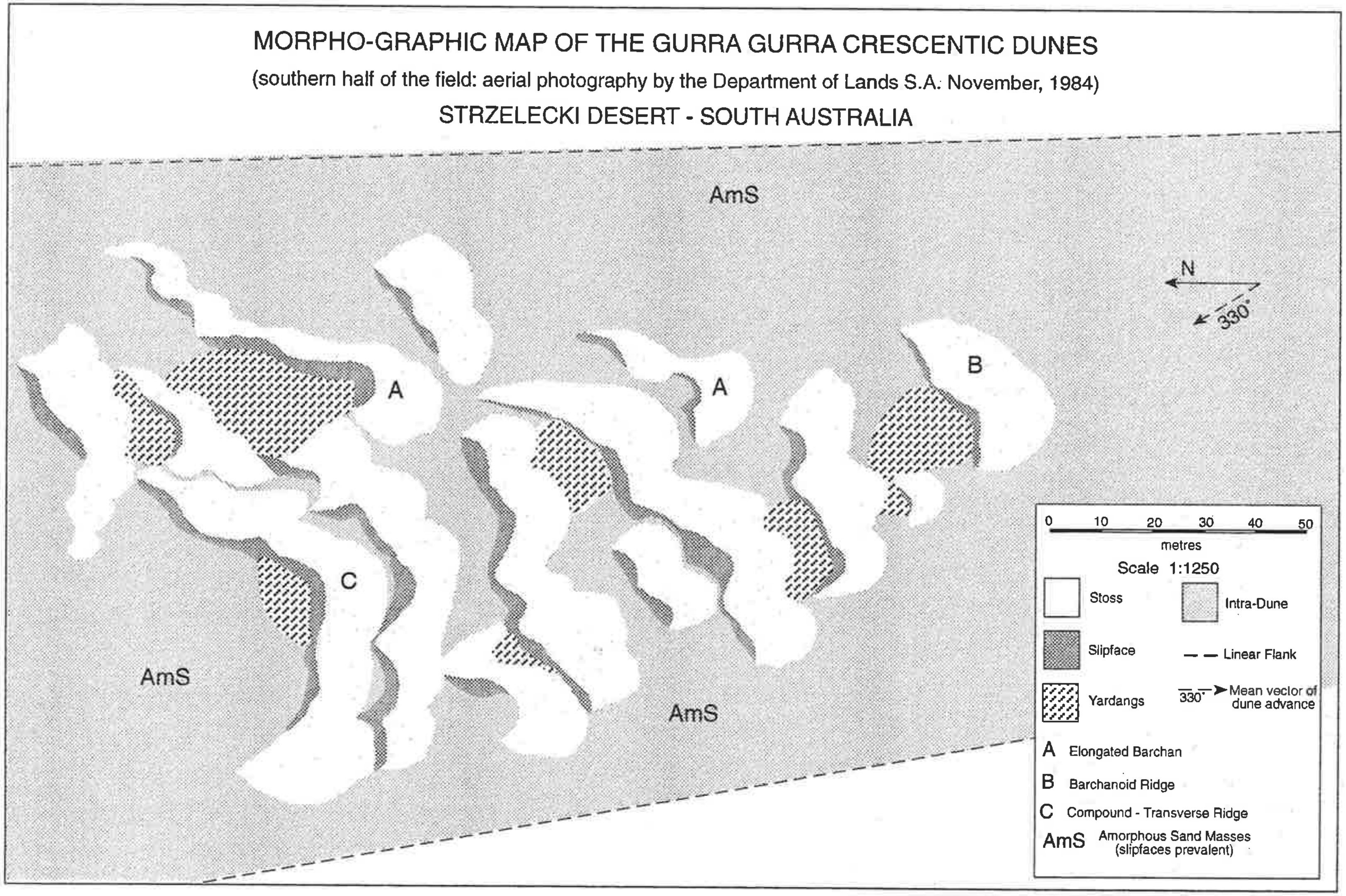


Figure 4.2 A morpho-graphic map of the Gurra Gurra crescentic dunes showing the southern and less evolved part of the dunefield in late spring - early summer, 1984. The northern sector of the field comprises only transverse forms. In the decade since, the dunefield has elongated north forming further transverse ridges. These dunes also appear more morphologically organised than in the suggested younger southern areas of the field. Note that yardangs are prevalent within the barchanoid-courts and that dunes (A) clearly demonstrate the genesis of a seif-linear form due to the elongation of the eastern horns.

dunes labelled A, Fig. 4.2). The longevity of the barchan-to-seif forms is short (< 10 years), as the inter-dune spacing and area of evolution is restricted to that of the underlying linear form. The development of a barchan dune reflects the strength and duration of the prevailing southern wind of summer. Hence, during the transition of crestal morphology from the coincident crest-brink of summer and the separated and reversed form of winter, the process of eastern horn elongation occurs. In other words, the uni-directional form of mid-summer, does not exist in the prevalent oblique air flow of spring. Similarly, the process of horn elongation diminishes as the gentler reversed winds of winter become the prevailing regime. Elongation may also be assisted in the transitory regime of autumn, when again the reversal of morphology and winds are gradually brought about by a phase of oblique westerly flow over the Strzelecki dunefield. The orientation of ripple patterns, scour flutes, and the presence of a secondary rear slipface on the eastern flanks of the spring dunes are indicative of stalwart westerly winds, while horn shortening is characteristic of the autumnal period.

In accordance with the model of Tsoar (1984), the Gurra Gurra barchans demonstrate a specific process-response mechanism driven by two dominant winds. Elongation is a response to these tandem wind directions, with the summer having greater strength than the slightly less effective vernal flow.

It is noteworthy that:

- only the horn opposite the secondary wind is elongated, a characteristic that differs from Bagnold's (1941) original model, but agrees with the model of Tsoar (1984).
- the angle between the orientation of the barchan ridge (an orthogonal line between the tips of the horns for a non-elongated barchan) and the alignment of the tail-end of the seif dune, is approximately 110° compared to 120° determined by Tsoar, and 080° for Bagnold, and is a vector sum of the summer and spring regimes, which themselves have a varying modal intensity depending on whether the season is waxing or waning.

Clearly, the dunes of Gurra Gurra waterhole identify with the 'barchan-to-seif' model of Tsoar (1984).

Ultimately, the final stage along the path of morphologic evolution is the compound or transverse ridge, which does not show the seif form, but rather elongates easterly with the barchanoid curvature of the dune ridge significantly diminished. The northernmost region, some 500 - 1000 m of the crescentic dunefield, is composed solely of well developed transverse dunes that straddle the entire width of the underlying linear dune and abut, but do not cross, the wide anabranching form of Strzelecki Creek. This area of the dunefield has considerably matured and expanded since the aerial reconnoitre of 1984 and from which Fig. 4.2 was produced. This suggests that the age of the crescentic dunefield is no more than about a decade, and proposes that crescentic development is a new phenomenon, or that phases of quiescence, destruction and reformation occur periodically, when climatic conditions significantly alter the micro-environmental responses of the area.

4.3 Crescentic Dune Morphometry

The following section investigates the interactions and mathematical relationships of dune morphometry using a comparative correlation and regression analysis of the Gurra Gurra (Strzelecki Desert, NE South Australia), Salton Sand Sea (Imperial Valley, California) and Pampa de la Joya (southern Peru) crescentic dunes, in order to discern site specific dimensional signatures.

4.3.1 Gurra Gurra Waterhole

Figures 4.3a-f illustrate the bivariate comparisons of dimensional parameters for the crescentic dunes of Gurra Gurra waterhole. Two data sets are used to mathematically characterise the changing but cyclical morphology of the Gurra Gurra dunes. Dune planimetry is derived from aerial images flown in late spring - early summer, November, 1984, and field measurements a decade later in autumn, (mid-April) 1993. The correlation

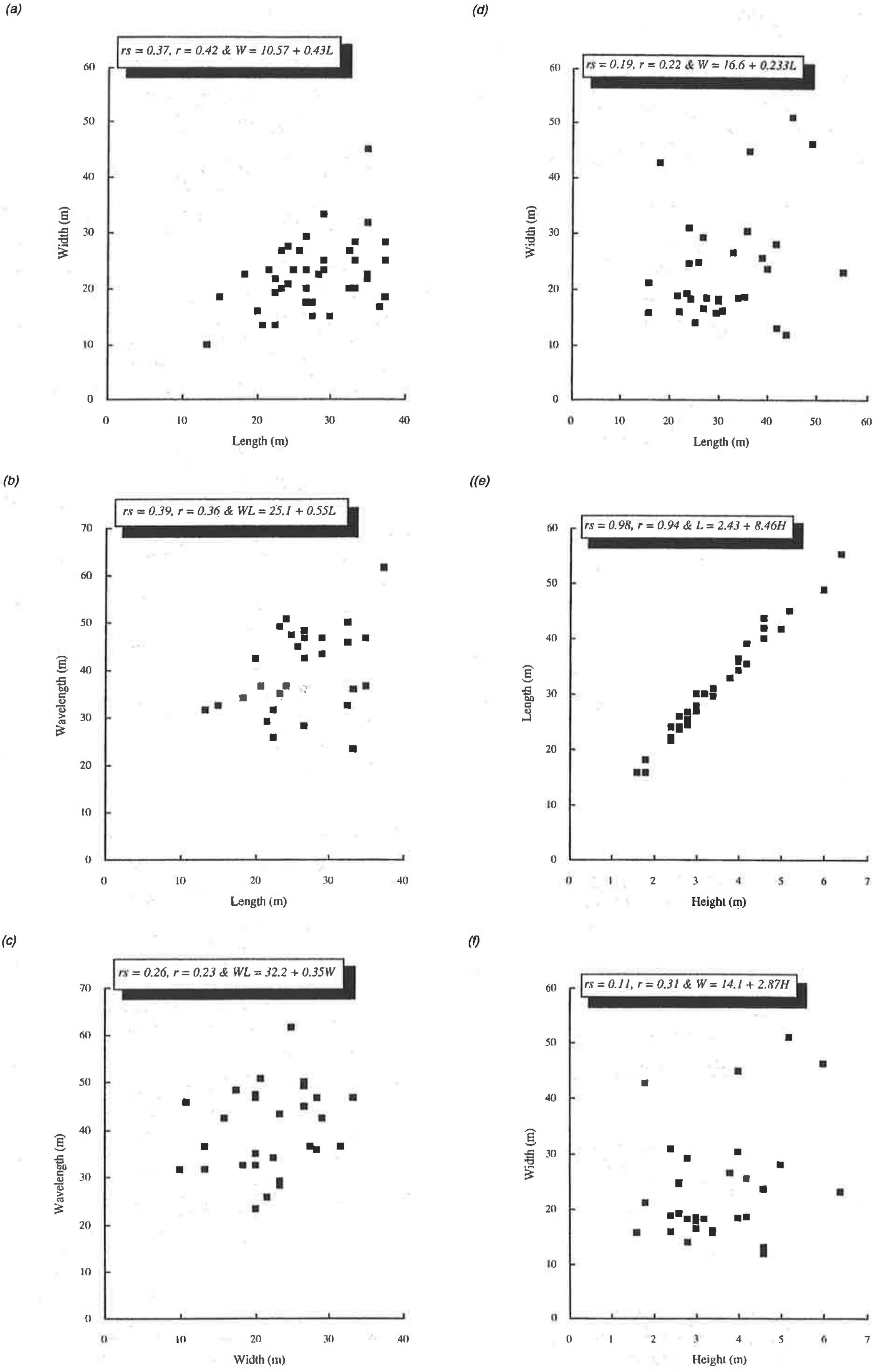


Figure 4.3(a- f) Bivariate scattergrams of length *(L)* width *(W)* and height *(H)* for the Gurra Gurra crescentic dunes, NE Strzelecki Desert, South Australia, as derived from aerial photographs flown in November 1984 (n = 41)and scattergrams of length *(L)* width *(W)* and height *(H)* as derived from field data collected in mid-April 1993 (n = 31).

coefficients ($r = 0.419$, $rs = 0.367$) for the relationship between width and length are low. However, they both are significant at the two tailed 0.05 level of significance.

Similarly, the F ratio for the regression equation,

$$W = 10.5 + 0.423L \tag{4.1}$$

is $F = 8.30$, and exceeds the critical F value. This rejects the null hypothesis *(H_0)* of 'no explanation' between bi-variables. There is, therefore, justification in accepting the alternative hypothesis *(H_1)* of the independent variable *(length)* being a reliable estimate of the dependent variable *(width)* at the 0.05 level of significance. The width of the Gurra Gurra dunes, in summer, is less than their length as generally defined by $W{\sim}0.4L$ or $L{\sim}2.5W$. Individual estimates of width using the regression should, however, be accompanied with a determination of confidence limits about that estimate.

Significance at the 0.05 level is narrowly ascertained between the wavelength and length of the summer dunes and is expressed by the regression equation:

$$WL = 25.1 + 0.550L \tag{4.2}$$

and an F ratio of $F = 4.36$. Dimensional correlation between wavelength and width is however insignificant and cannot justify the rejection of H_0 of 'no explanation' between these geometric parameters. Verification of this is given by an analysis of variance, where $F = 1.50$ relative to a critical value of 4.21 at the 0.05 level of significance.

The autumnal field data show no significant correlation between width and length. The corresponding F ratio of $F = 1.52$ cannot reject the H_0 of 'no explanation'. Thus, the regression equation (figure 4.2d) cannot be considered predictive.

Similarly, the testing of width and height for autumn also shows no significance of bivariate association at the 0.05 level of significance:

$$W = 14.1 + 2.87H \tag{4.3}$$

where $F = 2.98$ and $r = 0.305$, $rs = 0.114$.

Contrary to the absence of inter-dependency between width-length and width-height, there is a strong association of length and height. Both correlation coefficients are near complete ($r = 0.942$ & $rs = 0.978$) as is the F ratio at the 0.01 level of significance:

$$L = 2.43 + 8.46H \qquad (4.4)$$

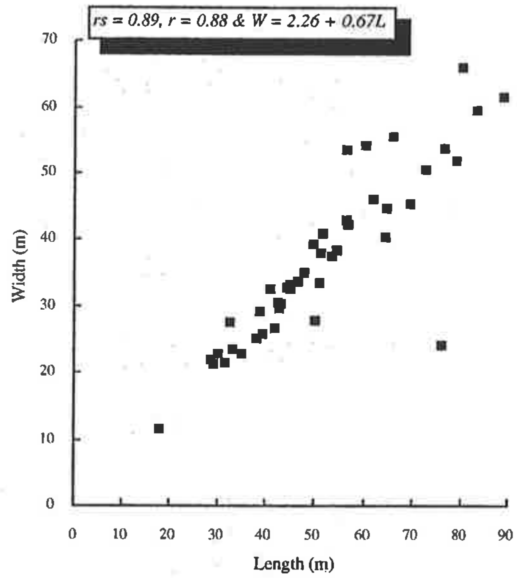
The test F value of 227.42 exceeds the critical F value of 7.56, thereby rejecting the null hypothesis of 'no explanation' between variables. Hence in autumn, the Gurra Gurra crescentic dunes generally indicate that $L \sim 8.5H$ or that $H \sim 0.12L$. Individual estimates of length using the regression should, however, be accompanied with a determination of confidence limits about that estimate.

Thus, several mathematical relationships exist between the morphometric parameters of the Gurra Gurra crescentic dunes. Both the absence and presence of bivariate correlation between geometric parameters identify the dimensional signature of the Gurra Gurra dunes, which in turn reflects the site specificity of the aeolian process. To demonstrate the inter-variable dependency and the specificity of locale on dune form, comparative analysis of bivariate associations for the same morphometric parameters has been applied to the crescentic dunes of the Pampa de la Joya and the Salton Sand Sea. The comparisons made between these different geographical localities, wind regimes, sand lithologies and physiographic settings, assist in understanding the evolution of the crescentic dune form.

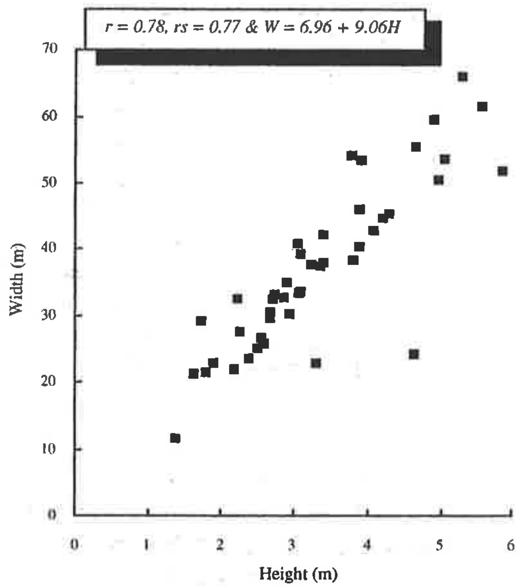
4.3.2 Pampa de la Joya, southern Peru

Using the field data of Finkel (1959) for forty-five barchan dunes of the Peruvian desert of Pampa de la Joya, highly significant correlations and regression lines were shown for the parameters of dune length and width. Strong correlations exist between width and length for all length variables at the two tailed 0.01 level of significance (Fig. 4.3). In like manner, the test F value exceeds the critical value, thereby rejecting the null hypothesis *(H_0)* of 'no explanation' and concludes that length may be used to make a reliable estimate of dune width from the regression equation.

(a)



(b)



(c)

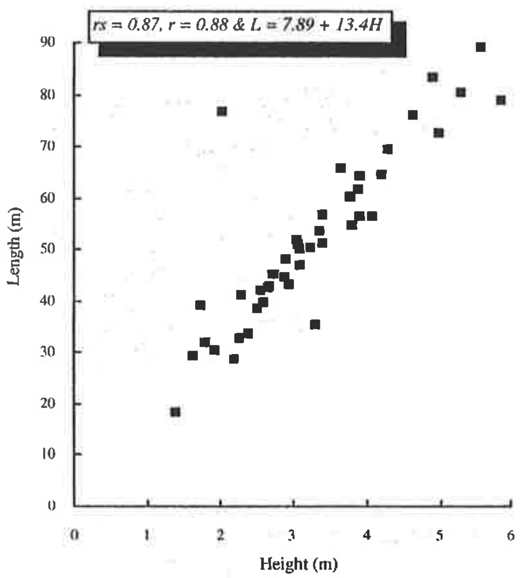


Figure 4.4 Bivariate scattergrams of length *(L)*, width *(W)* and height *(H)* for *(n=45)* crescentic dunes of the Pampa de la Joya region, Peru (data from Finkel, 1959).

$$W = 2.26 + 0.67L \quad (4.5)$$

$F = 138.74$ and the correlation coefficients are $r = 0.876$ and $rs = 0.892$. The dimensional ratio of length-to-width approximates 1:0.7 or W~0.67L or L~1.49W.

The Pampa de la Joya data has similarly given a strong correlation between dune width and slipface (average crestal) height[2]. A high estimate of correlation, $r = 0.782$, and an F-ratio of 65.95,

$$W = 6.96 + 9.06H \quad (4.6)$$

verifies an approximate height-to-width ratio of 1:9. Likewise, the expression of dune length versus dune height describes a strong relationship.

$$L = 7.89 + 13.4H \quad (4.7)$$

where $F = 138.42$ and the correlation coefficients are $r = 0.876$ and $rs = 0.874$.
The positive regression line illustrates how height increases in proportion to length with a ratio that approximates 1:13.5 for height-to-gross length.

Irrefutably, the bivariate association for the barchans of Peru do not differ from zero by chance, but portray highly significant dimensional causality between variables. The high degree of correlation between morphometric variables is suggestive of an equilibrated morphology.

4.3.3 Salton Sand Sea, California

The barchans of the Salton sand sea are approximately twice the size of the Peruvian dunes and offer an ideal scale comparison between like forms at different geographical localities. Using data from Long and Sharp (1964), the geometrical relationships for the barchans of the Salton Sand Sea also confirm the existence of dependence between dimensional elements. The Spearman correlation coefficient[3] for width versus length

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[2] Height of the Pampa de la Joya crescentic dunes is an average of the height of the crest above the toe of the leeward slope and the windward slope, as measured by Finkel (1959).

[3] It must be noted that non-parametric statistics are more pertinent for the data of the Salton Sand Sea, as the distributions of the morphometric variables are not considered to be significantly Guassian, although this may be due to a non-representative sample size.

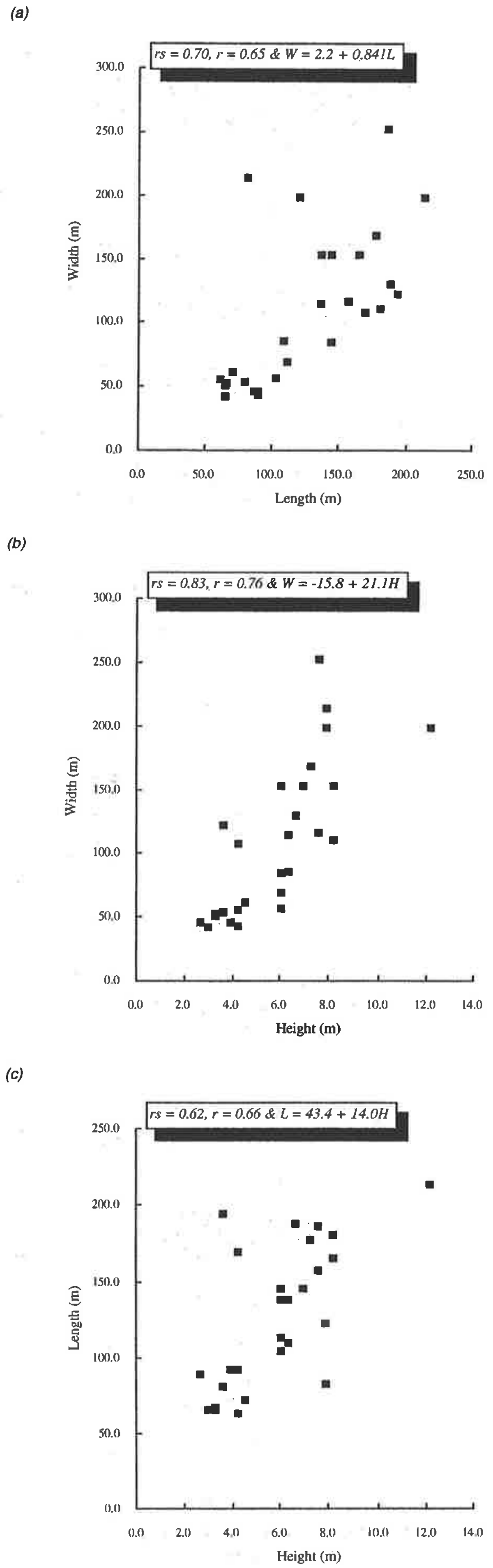

Figure 4.5 Bivariate scattergrams of length *(L)*, width *(W)* and height *(H)* for *(n = 27)* barchan dunes of the Salton Sea - Imperial Valley, California, USA (data from Long and Sharp, 1964).

($rs = 0.70$), width versus height ($rs = 0.83$) and length versus height ($rs = 0.62$) are significant at the two tailed 0.01 level of testing and show relatively strong positive relationships between bi-variables (Fig. 4.4). Due to the robust nature of the method of linear regression and despite the less than Guassian distribution of the data, it is evident that the Salton dunes also display highly significant expressions of prediction. Nevertheless, any estimates of *'y'* using the regression line should be accompanied by confidence limits about that estimate. Such expressions are, however, derived from data with a greater covariance and possess a lower degree of predictability. Dimensional signatures, therefore, reflect site specific influences on form, and the interaction of the two variables.

The regression equation for dune width versus dune length is:

$$W = 2.2 + 0.841L \tag{4.8}$$

where $F = 18.50$ and the correlation coefficients are $r = 0.652$ & $rs = 0.698$.

The linear expression identifies the geometric ratio of length-to-width as 1:0.84, $W{\sim}0.84L$ or $L{\sim}1.2W$.

The expression:

$$L = 43.4 + 14.0H \tag{4.9}$$

where both $F = 18.81$ and the correlation coefficients are $r = 0.65$ & $rs = 0.62$ are significant at the 0.01 level of significance for each test statistic, thereby rejecting a H_0 of 'no explanation', and confirming that dune height may be used to make a reliable estimate of dune length from the regression equation. The linear expression identifying the geometric ratio of gross length-to-height as 1:14.0, or $L{\sim}14.0H$.

All equations for the morphometry of the barchans of the Salton Sand Sea are seen to reject the null hypothesis of 'no explanation' between bi-variables and, therefore, favour the alternative hypothesis of the independent variable *(x)* being a reliable estimate of the dependent variable *(y)* at the 0.01 level of significance. The only expression that offers a substantial difference to the regression coefficient relative to the barchans of Peru, is the height-to-width ratio of 1:21

$$W = -15.8 + 21.1H \qquad (4.10)$$

where $F = 34.71$ and $r = 0.762$ & $rs = 0.826$ are significant at the 0.01 level.

4.4 Summary: Crescentic Dune Morphology and Morphometry

Dune morphology is a response to the environmental setting. Dunefields of like-dune type develop in settings that require a given number of like factors, irrespective of geographical location. However, each location also consists of site specific variability. Variability at the Gurra Gurra site includes:

- seasonal wind strengths and directions;
- the abundant supply of dune building sand;
- the incoherent substrate; and
- the obstruction to air flow caused by seasonal vegetation.

Dunes, therefore, adjust their form to the influence of environmental variables at a site. For example, the existence of influential reversing and oblique winds at Gurra Gurra waterhole confounds the establishment and maintenance of morpho-equilibrium. Compared with the prevailing uni-directional regimes and near-equilibrated dune shapes of the Pampa de la Joya and Salton sites, the Gurra Gurra dunes are very distant crescentic relations.

Site specific micro-meteorological conditions are not unusual in the Australian deserts, and are a major reason for the existence of crescentic dunes at Gurra Gurra in the Strzelecki Desert. Brookfield (1970) recorded high sand drift potential at Finke on the NW side of the Simpson Desert, a site that also has localised crescentic dunes, but a regional pattern of low sand drift potential. In accordance with both Fryberger (1979) and Brookfield (1970), the Strzelecki observations have shown the principal winds to emanate from the south and south-east during the hotter months, while during the cooler months, a higher percentage of northerly and south-westerly winds occur. The annual aeolian regime also varies with high speed winds during spring and summer, and low speed winds during autumn and the winter months. Despite there being a reversing wind pattern, the greatest sand drift potential results from southern winds in summer, followed by less dominant western and northern wind vectors in spring and winter respectively. This aeolian regime has a major influence on dune shape. The summer winds instigate crescentic dune formation, while the spring and winter winds develop the dunes into elongated and reversing forms. Spring winds have also assisted in developing some of the more widely spaced, individual crescentic dunes into a hybrid crescentic-linear morphology. Horn elongation has lead to juvenile seif-linear dunes, as previously reported by Bagnold (1941), Tsoar (1984) and Kar (1987, 1991). The reversing character of the dunes in winter, however, appears to interrupt this transformation, with the dunes becoming neither definitively crescentic or linear. Narrow dune spacing and coalescence appears to assist the termination of long term seif-linear evolution.

The morphologies of the crescentic dunes of Gurra Gurra waterhole are further explained and quantified by dimensional comparisons with the simple class of dunes on the Arequipa plain of Peru and the Salton Trough of California. In general, the Gurra Gurra compound dunes do not demonstrate strong bivariate relationships between different morphometric parameters. They also reveal an absence of dimensional correspondence with the simple dune forms of the Pampa de la Joya and the Salton Trough. The ratios of width-to-length and length-to-height for the Peruvian and Californian dunes, clearly demonstrate like-conditions of formation. Though even here, the regression equations, and a generally lesser degree of correlation between morphometric parameters, suggest the Salton Trough dunes have higher dimensional variability, as well as being much flatter than the dunes of Peru. It is clear that crescentic dunes exist as a spectrum of dimensional varieties within and across the boundaries of each major class of simple and compound types. Similarly, Breed and Grow (1979) concluded this for intermediate shapes found between the extremely curved and extremely straight varieties of compound dunes in the Algodones Dunes, California, the Peski Karakumy erg, U.S.S.R. and the Gran Desierto sand sea, Mexico. It was revealed by Breed and Grow (1979) that a large range of form and hence, dimensional correspondence, exists between intermediate varieties of compound dunes.

In this study, linear regression models have identified morphological variants, and that such shape variation is a derivative of the environmental conditions from which dunes develop. The more 'simple' the conditions, the less complex the dune form. The dunes of the Salton Sand Sea are in a state of simple-form quasi-equilibrium, while the barchan population of Pampa de la Joya approach an equilibrated shape. Comparatively, however, the crescentic dunes of Gurra Gurra waterhole are in a constant flux of dimensional change and varying states of compound-dune quasi-equilibria. Clearly, local site conditions, inclusive of geological, meteorological, biological and hydrological considerations, influence the genetic class of dune and, thereby, control the morphometry and morphology of dunes. For this reason, mathematical signatures of dune shape will demonstrate the similarity of the environments for dune development. Chapter 7 further explores the fundamental morpho-dynamic mechanisms and environmental setting which best explain the development of dune morphologies of the Gurra Gurra crescentic dunes.

In as much as morphometric differences are due to environmental conditions, the comparative mathematical impropriety of the Gurra Gurra dunes with those of the Arequipa plain and Salton Trough may also be exacerbated by their barchanoid-to-transverse nature being compared to the simple barchan form. Breed and Grow (1979) reported that like-dunes of compound barchanoid variety demonstrated very good to excellent dimensional correspondence to each other, but mostly poor to fair correspondence to transverse forms. This result is similar to the comparisons of the aforementioned sites. The compound form of the Gurra Gurra crescentic dunes does not correlate with the simple morphology of the dunes of Pampa de la Joya and the Salton Sand Sea. Compound form becomes less arcuate in plan-form and takes on the shape of a transverse ridge, whereas simple crescentic dunes such as the barchan and intermediate class, the barchanoid, have a higher degree of lee-side curvature. Air flow characteristics must also differ between simple and compound dune varieties, as each dune type projects a different shape into the boundary layer. Hence, the equilibrium shape is entirely different for the two classes of dune. Therefore, one cannot expect simple and compound classes to be similarly mathematically associated. Similar dimensional characteristics of the Gurra Gurra dunes may only be found with other barchanoid-to-transverse dunes.

Five fundamental statements of morphological and morphometric control are ascertained from this study.

These are:

- site-specific environmental conditions control the genetic class of dune;
- site-specific environmental conditions control dune shape;
- a spectrum of dune morphologies exist within and across the formal class boundaries of simple and compound;
- dimensional comparisons of similar dune-types reveal the degree of correspondence with shape-equilibrium; and
- the degree of dimensional correspondence between different dune classes is low.

5 Depositional and Erosional Landforms

5.1 Sedimentological Wind Vanes

The Gurra Gurra dunescape is rich in both depositional and erosional sedimentary landforms. Sharp (1963) and Howard (1977) have used surficial sedimentological features formed by air flow over and around dunes to successfully determine antecedent wind conditions. In like manner, the form and orientation of ripples, nebkha, lee projections, meso-yardangs and scour flutes have allowed the present author to deduce seasonal air flow regimes for Gurra Gurra waterhole.

5.1.1 Winter

Winter ripple-forms include in-phase catenary-to-linguoid and sinuous types, indicating a reversed air flow towards the south. Although some dunes demonstrate 3D asymmetrical catenary forms on the separated crest-brink zone of the dune, most dunes (especially the smaller 1-1.5 m high dunes) reveal the crest-brink and upper lee positions to have two dimensional symmetric straight transverse ripples, or very gentle sinuous transverse forms. The 2D forms typically show flattened crests and very low heights ($\leq$ 0.5 cm) with wavelengths between 5-10 cm ($10 < RI < 20$). The development of 2D form is most likely a consequence of a reversed wind flow from the north, where reactivation of the lee-ripple sands back over the ripple-crest returns the arcuate plan-form, developed in prior southerly winds, to a more symmetrical character. A paucity of ripple forms upon the mid stoss was found for many dunes, while the leeward slipface conspicuously has an absence of such features for dunes of barchanoid form. Those dunes with less barchanoid curvature of the slipface demonstrate pronounced lee slope ripple crestlines longitudinal to the direction of dune advance, with a wavelength of 9-10 cm and a height of 0.5 cm ($18 < RI < 20$) upon the mid-to-upper lee slope. Ripples are generally absent from the rear slipface. However, the leeward face of the dune linguoid does sometimes show minor ripples of low sinuosity and bifurcation migrating towards the east.

In contrast to the symmetrical transverse patterns upon the upper lee and crest-brink, the patterns of the upper horns and upper stoss elements are typically in-phase sinuous or catenary swept types with irregular or reticulate patches of out-of phase catenary ripples upon the mid-horn, returning to in-phase sinuous-to-catenary forms on the lower horn. The complexity of the patterns on the horns may be indicative of variable shear velocities over the changeable inclination of these elements and/or associated grain-size alteration. The upper regions of the dunes generally show a more coherent sinuosity where migration is to the SSE, while ripple orientation is towards the ESE for the lower horns. Horn gradients between the crest and interdune demonstrate a range of values between 0 °- 15° for the eastern horn and 1° - 20° for the western, with greatest inclination occurring between mid-to-lower horn slope. Generally, the gradient is < 10° over the majority of the element, with a median of 4° for horn-east and 8° for horn-west (refer to earlier Table 4.1). According to Howard (1977) this would indicate a deflection of *c.* 1.7 times the slope angle for the ripple normal and result in an approximate 26° deflection for a gradient maximum of 15° and a significant 38° degree deflection for a 20° inclination on the lower western horn. Likewise, the stoss slope with a median gradient of 8° would have *c.* 14° deflection, while the median lee inclination of 17° would deflect by as much as 32° on a non-avalanching slipface.

The development of ripples longitudinal to the primary flow direction along the base of the rear slipface may in part be due to divergence of northern flow extending over the secondary slipface and/or evidence of a secondary westerly wind vector. Either mechanism may explain elongation of the eastern horn and the subsequent development of some dunes into seif-linear forms. Although air flow divergence over the dune may assist in elongating the eastern horn, observations of easterly migrating ripple fronts caused by spasmodic but intense storm gusts from the NW, show how moistened sand in high velocity air currents can be stripped back by to a depth of some 1 cm - 3 cm by preferential drying of the surface laminae. Under such conditions, patterns of lunate individuals and in-phase catenary transverse ripple sets (height ≥ 1 cm) migrate towards the ESE upon the crest-brink, upper stoss and intra-dune corridors. Assuming these morphologic elements to be sub-horizontal, the normal to the ripple crestline can be considered parallel to the regional wind pattern. Ripple patterns formed by antecedent winds are readily erased under such conditions. The presence of a dominant easterly oriented pattern is presumed to be formed rapidly (within minutes) but may persist for days to weeks until more persistent but gentle reversed northerly winds overprint this easterly pattern. It is here suggested that this process explains the development of interference ripple patterns observed on the winter dunes. Interference or overprinting ripple patterns are most prevalent on lower stoss slopes where both transverse (secondary pattern) and longitudinal (primary) patterns are observed each with a wavelength of 2-7 cm and 10-15 cm respectively, and with amplitudes not exceeding 0.5-1 cm. ($7 \leq RI < 30$). Further examples of oblique interference patterns occur on some eastern dune horns, where ripples of very slight sinuosity and bifurcated nature occur.

Wind sculptured features in more cohesive sands of the crescentic dunes exhibit southerly oriented scours, 0.1-0.2 m in dimension. Likewise, some meso-yardangs nestled between dune horns and consequently protected from easterly winds, display blunted leeward prows or definite bi-versal forms. These are indicative of reversing and more gentle northerly winds of winter, relative to the formative and dominant summer southerly air stream. The protection afforded these features, which are usually located in the courts of barchan and transverse dunes, may be one reason for their survival. The existence and preservation of these features in the presence of the reversed winter wind flow identify the low intensity of this regime, especially within the crescentic intra-dune corridor. If the winter winds were of significant saltation strength and duration, then destruction of the loosely, cohesive meso-yardangs would occur, as the narrow leeward slope would quickly abrade under the influence of higher velocity, near-surface winds. Observations by the present author of daily winds show that saltation periods are few and concentrated mainly in the early afternoon hours. In contrast, periods of quiescence dominate. These observations are similar to those reported by Brookfield (1970) from the southern Simpson Desert. The existence of these fragile features throughout several seasons, as well as the blunted form of the leeward slope, scour features at the base and the orientation of centimetre-size scour flutes, strongly signifies the low intensity and reversed nature of the winter winds. This provides evidence for a repetitive cycle of wind reversal that extends beyond the seasons for which data was collected. The presence of sedimentary features as well as this author's direct observation of winter saltation in a southerly direction, further indicates the reversed aeolian regime. Sinuous-to-catenary transverse ripples within the intra-dune corridors of both the yardang fields and crescentic dunes, and minor scour flutes on the northern side of these obstacles, also clearly demonstrate a predominantly northern origin for the primary winter air stream.

5.1.2 Summer

The wind pattern of summer is converse to that of winter. During summer micro-landforms demonstrate a predominant southerly air flow. Sharply defined and sinuously-crested lee projections extend from some barchanoid elements of several dunes exemplifying a northward transport direction (Fig. 5.1). Similarly, sinuous transverse ripples (wavelengths ~13 cm and amplitude ~0.5 cm, *RI* ~ 26) dominate the coincident crest-brink of larger dunes and further identify the ripple normal to be northerly (330° - 340°) and in the direction of dune advance. As during winter, the larger dunes display a plethora of ripple forms and orientations in response to the variable inclination of the horn slope, grain-size and wind speeds that occur between the interdune and crest. Straight to sinuous transverse and in-phase catenary-lunate are the dominant micro-forms occurring on the horn between intradune and crest, with the less sinuous patterns found in the lower morphological zones of suspected lowest shear stress (Fig. 5.1). Straight ripples develop upon leeward slopes that are not under active avalanche. Ripples are most prevalent upon the linguoid elements which possesses a gradient less than the angle of repose and least on the 32° inclination of the barchanoid.

The relatively low shear stress of the wind upon the interdune produces ripples that are generally less sinuous than those found upon the crest-brink. This infers that relative to a

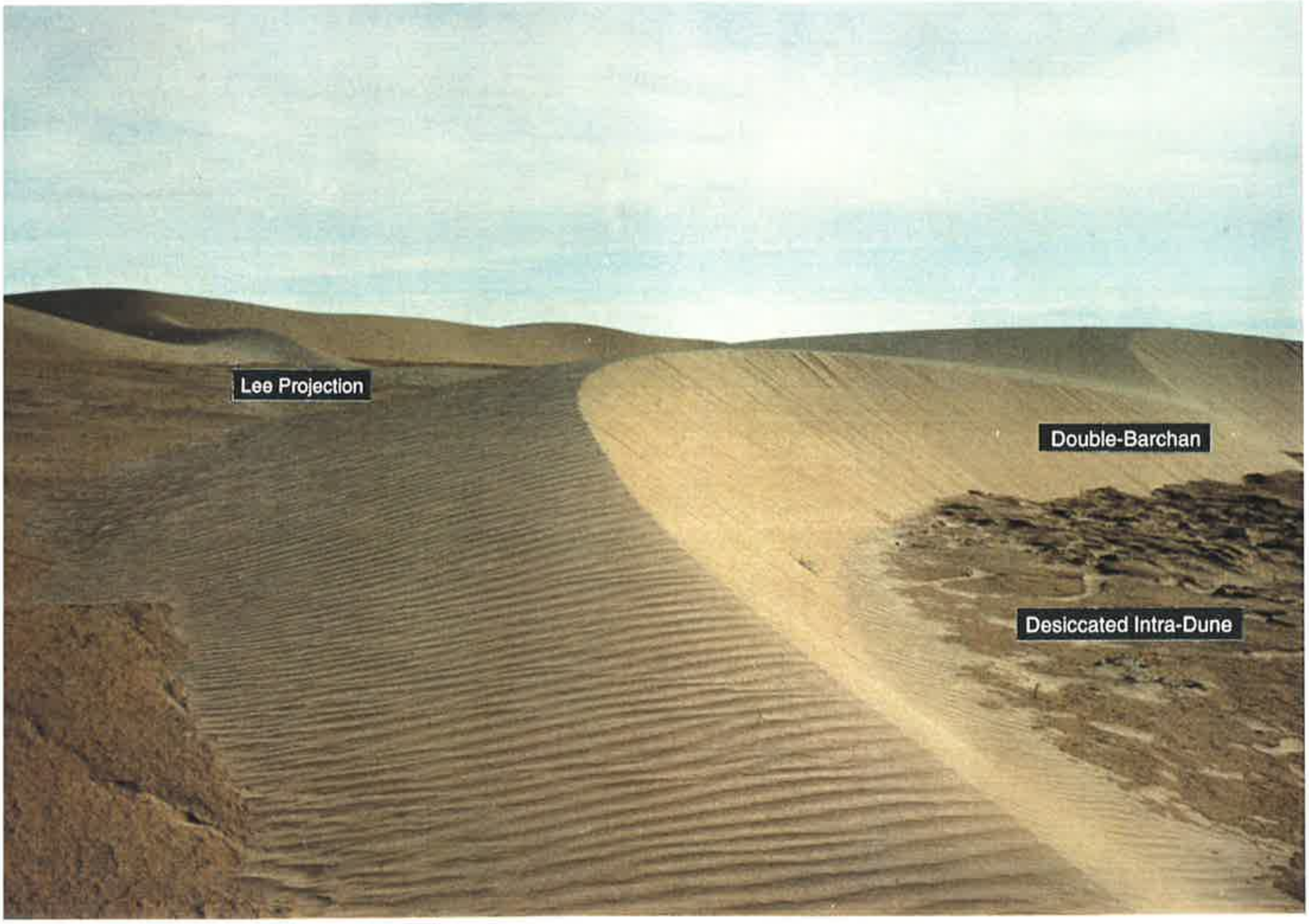

Figure 5.1 The summer intradune shows the residence of heavily eroded yardangs within the northerly facing barchanoid of a double barchan. The less desiccated character of the intradune between dune ridges, results from airflow that is less vehement in the downcutting of the eluviated montmorillinite rich laminae of the quartzose intradune floor. Significant erosion of the polygon structure is so far absent at this locality. The sharply delineated coincident linear-linear morphology of the summer dunes is also displayed with the double-form (~ 2 metres high). Except for the dunes themselves, the intradune is relatively free of loose sand patches. Slipface avalanche deposits are plentiful, whilst ripples of sinuous in-phase form with minor bifurcation migrate over the stoss and horns. However, leeward aprons are absent. Such features infer transport of the total dune mass towards the NNW. These sedimentological features are in contrast to those formed in the other seasons when reverse and/or oblique winds are operative. The background clearly shows a lee projection (~ 2 metres high) extending from the barchanoid element of the dune ridge. This symmetrical structure clearly identifies a northerly direction of air flow and is a feature only found in summer. The horn of a nearby dune ridge interacts with the projection on the left hand side, and has assisted in enlarging the dimensions of the feature at the point of mergence.

dune's crest, an increase in upslope velocity may increase the sinuosity, although differences in grain-size may also have a significant influence here. Straight to gently bifurcating sinuous patterns longitudinal to the southerly vector of air flow occur on smaller dunes at both basal and upper morphologic positions. This little altered ripple form indicates either an absence of a significant increase in shear velocity over smaller structures (≤ 2 m high), for example dunes 3, 4 and 5, or that grain-size variation over the stoss of the dune is not significantly different. Abundant downward longitudinal fluting patterns, that resemble indistinct and ill-developed ripples, occur along some dune slipfaces, especially those with a low level of concavity in plan-form and suggest the influence of minor oblique winds. Ripple-less surfaces occur over the concave portion of the lee, albeit, sinuous and longitudinal bifurcating forms develop on the linguoids of the lee slopes as in winter. The easterly orientation of the ripple normal is too great to be attributed solely to the deflection of the primary southern wind vector by slope gradient and is considered as evidence of a secondary wind from the west. Interference patterns between the primary southern and the westerly secondary vectors were not discerned on any part of the dune. This observation, however, does not correlate with the Oodnadatta sand roses, where more ESE-to-E vectors are prevalent in January-March. It is likely that local conditions are operative within this smaller component of the larger Simpson-Strzelecki dunefield and/or the difficult task of discerning wind direction for near symmetrical, bifurcating ripples. Nevertheless, both dune and intradune ripple patterns reflect origins under the influence of a dominant southern (*ca.* 160°) vector. Fluting upon the underlying aeolinite (Fig. 5.2), as well as erosion scours around obstacles and the leeward orientation of the abundant 1.5 - 3 m high nebkha of the interdune region (Fig. 5.3), also support this conclusion. However, a secondary and weaker westerly vector is intermittently active, and may assist in eastward dune elongation. The high shear velocities of summer generate ripples of greater sinuosity than do the lower shear velocities of winter.

Figure 5.2 U-shaped erosion moats (flutes, ~0.5 metre deep) circumventing an obstacle of weakly clay-cemented sands (relict aeolinite) that has been stabilised by vegetation upon the lower western flank of source linear dune. The abrasion patterns afforded by flow vortices around such obstacles, aptly describe both regional and local wind directions. Such wind orientations are also depicted by the innumerable 'sand shadows' behind dead grass and bush clumps upon the interdune corridor. The general appearance of a loose sand-saturated environment is typical of summer conditions throughout the region.

Figure 5.3 Nebkha some 1.5 to 3 m high upon the interdune of the Gurra Gurra site. The tapered, streamlined form of the leeward sand deposits are ideal wind vanes. Where such landforms occur, sand-free corridors are prevalent and are indicative of high wind velocities and smooth uninterrupted local topography.

5.1.3 Autumn

Autumn ripple patterns are predominantly sinuous-transverse and catenary-transverse upon the upper morphologic positions of the dunes, expressed in ripple indices between 7 - 30, as found in the other seasons. Ripple migration is northwards and correlates with the prevalent coincident crest-brink morphology of the gross form and the southerly wind vector inherited from the previous summer. Minor sinuous, longitudinal-bifurcating ripple forms are found along the lee slope, being most prevalent upon the linguoid elements of the transverse ridges, while the barchanoid of the ridges is mostly devoid of ripples. Most significant is the random, but rare, occurrence of landforms that signify oblique wind action. Sculptures that resemble micro-yardangs reveal an easterly wind vector across the dune-field. These micro-sculptures are extremely fragile forms eroded from the intradune and measure between 10 - 20 cm in height and < 10 cm wide. They display some fundamental streamlining, although their fragility and size does not allow the typical yardang shape to evolve. The lower windward face displays undercutting and the upper windward surface has 'finger-like' projections jutting windward. The form tapers leeward with greater width along its length for the upper one-third, indicating greater abrasion and/or deflation along the lower two-thirds. Streamlining is definitively westward and identifies the existence of a minor oblique wind vector of easterly origin. The symmetrically tapered lee shows the convergence of streamlines, while fluting and a peripheral moat also depicted wind direction and the intensity of vorticity-abrasion along the intradune floor. The process of differential erosion is clearly seen to occur where clay enriched-cemented layers, often only some 2 mm thick, extend as ledges along the length of the structure. It is without doubt that such micro-structures demonstrate the erosive action that downcuts the intradune floor to produce larger yardang fields from the sculpturing of smectite-enriched desiccation polygons. This, in turn, aids in the supply of sediment to developing crescentic dunes downwind. Whitney (1983, p. 236-37) also identified very similar sculpturing of yardangs at the meso-scale in the Kharga depression, Egypt, and pointed out that the protrusion of delicate 'finger-like' projections from the windward surface must indicate the "retardation of erosion at centres of vorticity" most probably where reverse and normal flow unite.

5.1.4 Spring

The ripple patterns of spring are sinuous-in-phase-catenary over the stoss with migration towards the south. The stoss slopes of the spring dunes clearly demonstrate the orthogonal interference pattern between northern and western wind vectors (Fig. 5.4), with easterly migration suggesting an overprinting of the primary southern orientation. This observation coincides with the diurnal pattern of winds, where northerly saltating winds consistently develop between 1130-1230 hours, and intensify and surreptitiously shift towards the west in the late afternoon. The NNE migration of coexisting out-of-phase catenary-to-lunate ripples and in-phase sinuous-transverse ripples, together demonstrate the influences of a variable stoss gradient both across and over the dune, variable grain-size and diurnal wind pattern. Longitudinal lee ripples are found to have wavelengths between 8-15 cm and amplitudes of ~0.5 cm ($16 < RI < 30$) in the majority of instances. Several dunes displayed upper lee transverse sets that are significantly smaller, with wavelengths of 5-8 cm and amplitudes ≤ 1 cm, and which also corresponded to ripple dimensions and orientations upon the crestal zone. The stoss slopes demonstrate larger ripple sets of wavelengths 15 - 25 cm with amplitudes of 1-2 cm ($7 < RI < 25$). The larger the wavelength the greater the ripple height. Ripple indices ranging between 7-25 are not uncommon. Similarly, coarse ripples of the intradune floor produce indices of 17-20, while easterly migrating granule ripples upon the floor of Strzelecki Creek have wavelengths of ~50 cm and amplitudes of 5 cm ($RI = 10$).

Reversed flow is also recognised in spring by saltating sand streaming off the separated crestline of the rear slipface. The mid-to-low stoss slope and horns also display reversed flow for the sinuous ripples, with the sinuosity becoming less as the slope gradient and shear velocity decreased closer to the intradune, irrespective of any granulometric change. The ripples of the sub-horizontal crest-brink region (l~8-10 cm and amplitude of 0.5-1 cm : $16 < RI < 20$) reveal a low degree of sinuosity. Minor straight-to-sinuous, bifurcating, leeward ripples migrate eastward and once again occur on the linguoid of both barchanoid and transverse dune types. The presence of a second reversed slipface, oriented normal

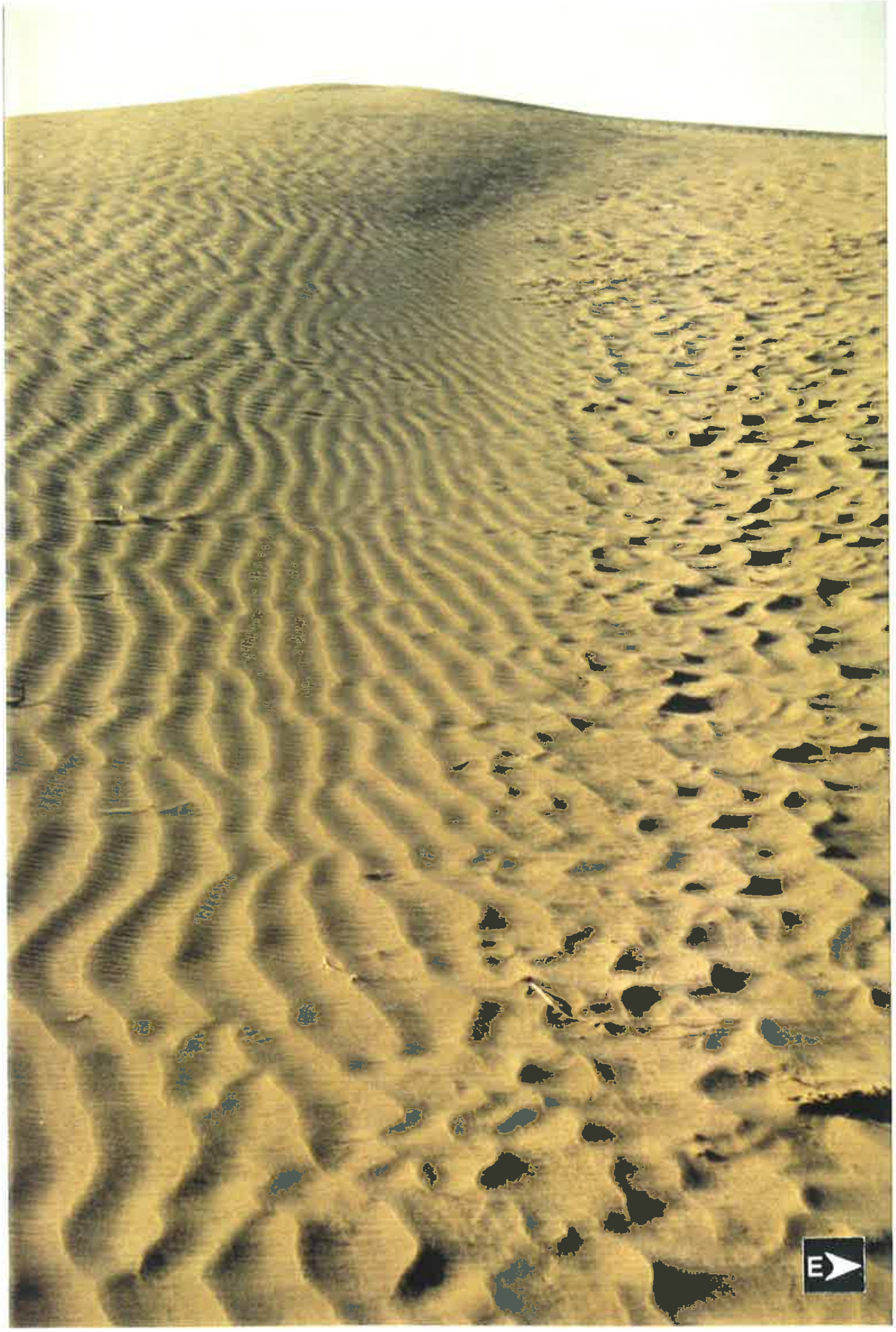

Figure 5.4 Changing ripple patterns upon the stoss slope of a spring dune. The variation between continuous near symmetric sinuous (wavelength ~ 15 - 25 cm, height ~ 1-2 cm) and discontinuous catenary in-phase forms are sharply delimited by change in gradient and grain-size. Each ripple type characterises a dominant eastward direction (towards the right) of migration for the primary ripple set, while an interference pattern of orthogonal secondary ripples is evident along the troughs of the primary forms. This identifies a southern migration under the influence of a reversed wind in the morning hours of each day. The morphology of the primary ripple is a result of this diurnal transition of wind direction and is transient between asymmetrical and symmetrical. The darker albedo of the troughs is due to the accumulation of heavy mineral fine grade sands (2ϕ - 4ϕ) that consist of ferruginous lithic fragments, tourmaline, plagioclase and minor zircon and rutile.

to the east, as well as easterly migrating interdune catenary-transverse ripples, exemplify the presence of an effective western sand-transporting wind vector. Nebkha with partially circumventing scour moats on the western and northern perimeter also support this conclusion.

5.2 Slope Inclination and Ripple Orientation

The deflection of ripple patterns is undoubtedly assisted by slope inclination and gravitational creep. Slope angle also influences the divergence-convergence of air flow resulting in variability of shear velocity, a finding that is in accordance with the work of Howard (1977) (Section 2.13.3). The localities most suited for measuring ripple patterns as regional wind sensors are those that lie upon relatively open, unobstructed terrains such as the interdune, intradune, crest and upper stoss areas. Here, the terrain is fundamentally horizontal and has minimal impact on the orientation of the ripple pattern. Any other morphologic components of the crescentic form will offer variability to the orientation of ripple pattern and, most probably, ripple form.

The median gradient for the lee slope in winter is 17°, 32° in both summer and autumn and 22° for spring, while the median inclination of the stoss slope is between 6° - 8° for all seasons. The grade upon the horns averages < 10° for all seasons. However, the range shows that higher values commonly occurred about the mid-to-lower morphologic positions of the dunes (see earlier, Table 4.1). In all cases where ripples formed, it is necessary to understand that ripple orientation is not transverse to the regional (seasonal) wind regime, but is affected by the gradient of the dune slopes. Howard (1977) determined that for slopes of < 10° and for slopes approximating 20°, the quantity of deflection is between 1.7 to 1.9 times the slope gradient, respectively. Hence, for even the lesser grade of the stoss element, a deviation from the regional flow direction can be as much as 14°, whereas upon the non-avalanching winter lee, flow deflection could be as much as 32°, a significant cardinal variation. For the winter dunes, this quantity of variation may almost be significant enough to account for the orientation of ripples upon the lee slope linguoids, without the necessity of invoking the presence of a secondary easterly-oriented flow pattern. Without the observations of intradune, crestal patterns and scour flutes around nebkha and stabilised mounds, it would be justified to assume that flow deflection is the sole cause of the ripple orientations upon the linguoids. Caution must therefore be exercised in the examination and determination of air flow using ripple patterns. Alone, such patterns are unreliable, except for those on a horizontal plane. Future studies need to examine the influence of this effect in detail, as not only does gradient vary, but also the grain-size distributions and flow velocity characteristics for different slope elements. Change in each parameter alone, will cause variation to ripple morphology and orientation.

5.3 Depositional Landforms: Lee Projections and Nebkha

As postulated for the Thar Desert of India-Pakistan by Kar (1987), the Gurra Gurra dunes demonstrate some accordance of linear dune genesis from barchan dunes, where sandy cones extend from the lee slipface to join the adjacent ridge. Many morphological features of the barchanoid dunes, in both the Thar and Strzelecki localities, are similar and they most probably derive from similar aeolian mechanisms. Although the union of the barchanoid dunes at Gurra Gurra waterhole produces a morphology not unlike that of a linear dune, the form is still typically transverse to the principal wind direction, with a discernible slipface and stoss slope. Nevertheless, the curvature of the barchanoid is at a minimum.

Structural elements of the crescentic dunes at Gurra Gurra waterhole that lend credence to the barchan-to-linear hypothesis of Kar (1987), are lee projections (sandy cones) emanating from the barchanoid of the dunes during summer. A very conspicuous example of a major lee dune emanating from the barchanoid element of a dune is shown in Fig. 5.1. This feature is a linear symmetrical extension some 20 m long, 2 m wide and *c.* 1.5 m high, forming upon a relatively 'smooth' and consolidated intra-dune floor. The transverse profile consists of two gently concave slopes converging to a sharp straight crestline. In-phase transverse sinuous ripples extend parallel to the form's length, coinciding with a transport regime towards 330°± 5°. This feature was not present, or in any way developing, during the previous winter and is considered a consequence of the south-easterly summer winds converging and decelerating leeward of the barchanoid ridge crestline. Flow separation of high velocity, sand-laden winds, combined with an apparent leeward convergence of flow, results in sudden velocity deceleration and a gradient contrast that results in leeward deposition, concentrated normal to the slipface. The streamlined form of the lee projection may be assisted by vortices of flow around the perimeter of the dune as well as along the projection itself. As stated by Lancaster (1993), deposition dominates where air flow crosses the crestline at angles approaching 90°, a characteristic prevalent for the summer dunes of Gurra Gurra waterhole. The strong symmetry of the lee projection indicates that concurrent oblique flow-lines (vortices?) stem from the obstruction of the crescentic dune, that these flow lines are of approximate equal intensity and size, and that they are necessary for lee dune growth and elongation. Similarly, less well developed lee projections or sand cones were also found on many other dunes. Analogues of these features have also been reported by Cholnoky (1902), Olsen-Seffer (1908), Beachell (1910), Cornish (1914), Bourcart (1928), Cooper (1958), Norris and Norris (1961), Reid (1985) Wopfner and Twidale (1988) and Kar (1987, 1990) in other ergs around the globe.

At the transverse end of the field, where the wavelength between dune ridges has decreased, and a stacking of the transverse ridges is evident, dune elongation occurs by ridge extension onto the stoss toe of the dune in front. Bagnold (1965, p.214) used the term 'fulji' to describe these features of interconnecting dunes and intervening hollows. Fulji were prevalent during the reversing-oblique winds of spring at Gurra Gurra, and further demonstrated the complex dune morphology of these multi-directional, evolving crescentic dunes. This data further supports Kar's (1987) observation of crescentic ridge unification, leading to infilling of intervening voids and the production of a linear dune-form. In several areas of the transverse dunes of Gurra Gurra waterhole, this process was observed, but not extensively. In all other seasons, lee projections were absent and the dunes were essentially individual barchanoid or transverse ridges, lacking the lee-to-toe coalescence of the summer form. It is suspected that the reversing winds of winter assist in readjusting the form to a less coalesced and more widely spaced morphology. For dunes that have undergone significant lee-to-toe mergence, as with the reversing transverse forms of spring, distinct inter-ridge cavities occur and persist.

In summer, the accretion of lee dunes proper occurs behind obstacles to airflow, such as 'aeolinite-like' mounds often stabilised by dead shrub (Fig. 5.2 and 5.3). Significantly, such dunes are absent from the lee of yardangs, with the yardang groups enhancing deflation and abrasion around the obstacle, thereby limiting deposition and the formation of lee projections and dunes. The presence of yardang courts is assumed to cause turbulent leeward flow lines coming off and around the dune. This results in piece-meal convergence with deposition into a streamlined leeward projection being inhibited. Greeley (1978; 1986) identified a 23% increase in u_* in the lee wake of the Amboy cinder cone, relative to the ambient level of surface shear stress. Hence, lee dunes are prominent only within the open, relatively unobstructed intradune and adjacent interdune areas of the region. Lee deposition is encouraged by the obstacle shape and local wind conditions.

Lee deposits range in size from a few cm to > 10 m in length, the larger taking the form of streamlined nebkha (Fig. 5.3). For all lee dunes the leeward ridge is symmetrical and linear in profile, with a concave taper in the direction of air flow. The ridge line forms a sharp apex along its entirety with in-phase transverse sinuous ripples either side. Both ripple and lee-dune orientations are accurate determinants of wind direction. Where surface sediment is unconsolidated or more erodible, erosion scours, moats or current crescents circumvent the obstacle and lee-dune combination. Scours are typically U-shaped furrows (Fig. 5.2) which may extend the entire length of the obstacle, sited in relatively unconsolidated sediment. They are formed by a horse-shoe vortex of flow around a bluff body projecting into the boundary layer. The scours are capable of penetrating to a depth of ≤ 0.5 m and maintain perfect symmetry in plan-form and a distance of *c.* 20-30 cm from the perimeter of the obstacle. As with depositional sedimentary features, erosion marks are also definitive indicators of current direction.

5.4 Erosional Landforms: Yardangs and Desiccation Features

Yardangs are major erosional meso-landforms of the Strzelecki-Gurra Gurra crescentic dunefield. The profile of the Strzelecki meso-forms shows a strongly asymmetric morphology characterised by a steep windward prow and a gentle leeward slope (Fig. 5.5 and 5.6). Similar morphology was observed for the meso-yardangs of Peru by McCauley *et al.* (1977), who also inferred that changes in the length-width ratio are indicators of age. The morphometry of many of the larger yardangs (0.5 m x 5 m) at Gurra Gurra, gave a width-to-length ratio of 1 : 10 (n = 10). The erosional forms of this dunefield showed characteristics of forms observed elsewhere, and like dune-forms, showed that geometrical similarity also infers dynamic similarity in genesis and development. The Strzelecki meso-yardangs typically develop in groups, where wind flow erodes the relict linear dunes and sculptures the quartzose aeolinite into both flat-topped and tapering hull-shaped forms that typically expose the low angle cross bedding of the original linear dune (Fig. 5.6). For many duneforms, micro-yardang groups are prevalent within the courts of the lee-slipface. This observation is illustrated in Figures 4.2 and 5.1 and is in agreement with observations of suggested lee-side erosion of transverse dunes in eastern Mauritania by Severnet (1943) and Monod (1958).

At the southern and less transverse end of the crescentic dune field, groups of yardangs are not currently associated with dunes. The dunes have migrated forward, abandoning the yardang group, or the yardangs formed independently of the dunes. The prevalence of yardang groups within dune courts is evidence of landform persistence despite the erosive actions of multi-directional winds. The dune horns and their leeward curvature protect the yardangs from oblique winds. Where yardangs have greater exposure to multi-directional currents, either through dune advance or dune re-configuration, the form undergoes re-orientation and streamlining under a complex wind pattern. This results in a non-streamlined, jagged and generally non-descript structure. At best the form is one of a knob or pinnacle, resulting from erosion from all directions at different times of the year. As commented by Whitney (1983, p. 244) "...an aerodynamic system is far more complex than

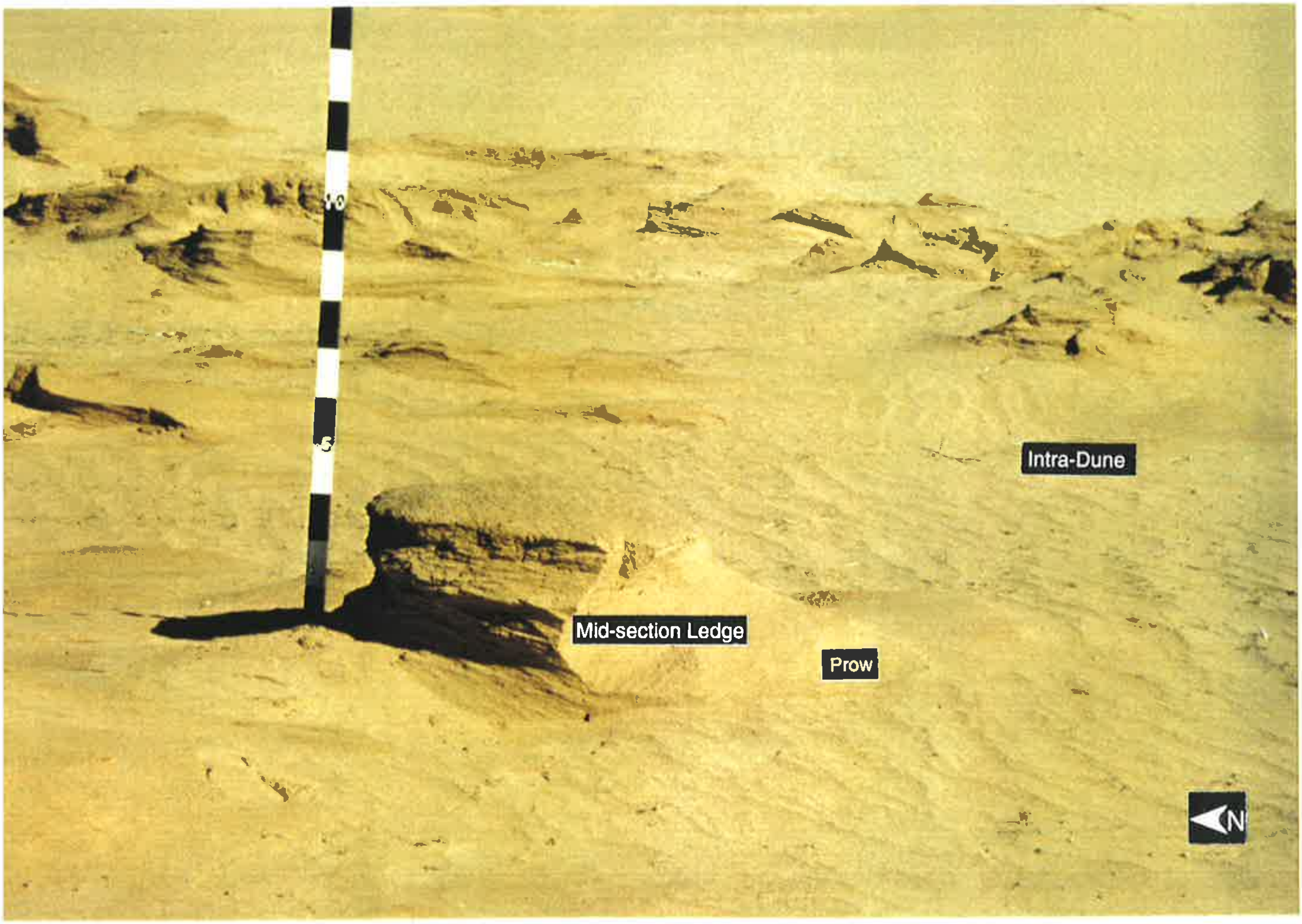

Figure 5.5 A well-developed flat-topped yardang cut into the weakly indurated relict linear dune some 14,000-25,000 years old. Such landforms are characteristic of the southern (less mature) end of the crescentic dune-field. Both abrasive and deflative sculpturing of form are responsible for longitudinal windward tapering. The leeward prow on the right hand side shows evidence of backcutting through abrasion, with the convergence of more gentle flow vortices windward, resulting in an absence of ripples and loose sand patches in the wake of the obstacle. While basal scouring is pronounced, two distinct and separated flow patterns are shown above and below a mid-section ledge that indicates an area of minimum erosion. Similarly, the flat, clay-rich upper surface marks the position of a prior intradune floor and attests to negligible erosion over the top of the feature.

many investigators have realised and varies within every alteration in shape and detail in a multi-directional wind, a different aerodynamic system results on any one feature with every wind shift, because the relationship of wind direction to feature orientation is altered and so alters the division of the streamlines and the subsidiary patterns."

Two yardang morphologies are prevalent amongst the Gurra Gurra dunes, these being, flat-topped and tapering forms. The flat-topped form is located in groups that reside at the southern-most end of the crescentic dunefield in areas outside of the protection of the courts of the dune barchanoid. The widest component of the meso-yardang is the windward side that is aerodynamically shaped to an apex. In plan-form the morphology resembles equilateral and isosceles triangles merged at their bases. Scour moats are typical at the prow of the equilateral form and terminate at the corners of the base angles. The alignment of these features is parallel to the major north-westerly direction (320°±10°) of dune advance, with the leeward section narrowing in this direction. The average dimensions of each individual amongst the meso-yardang groups are not much greater n size, if any, than the dimensions of the primary desiccation polygons of the intra-dune floor. This may infer a genetic relationship between the polygon and yardang (Fig. 5.7). Such an association may be due to preferential scouring along the fracture fissures around each polygon, thereby limiting the size of the yardang eventually formed from erosion of the intradune floor. The upper surface is preserved and remains horizontal, is clay capped and represents the previous intradune floor. This landform is strong evidence of downcutting of the intradune floor with the erosional sculpturing of yardangs resulting from this process. The longitudinal profile depicts an undercut prow or windward nose, with the upper prow also eroding backward, and developing a pointed windward face. The mid-section of the yardang is the widest portion of the yardang, indicating zones of preferential erosion along the length of the form. Undercutting by some 5 cm is also preferential along and below the flat top (refer to and compare Figures 5.5, 5.6 and 5.7).

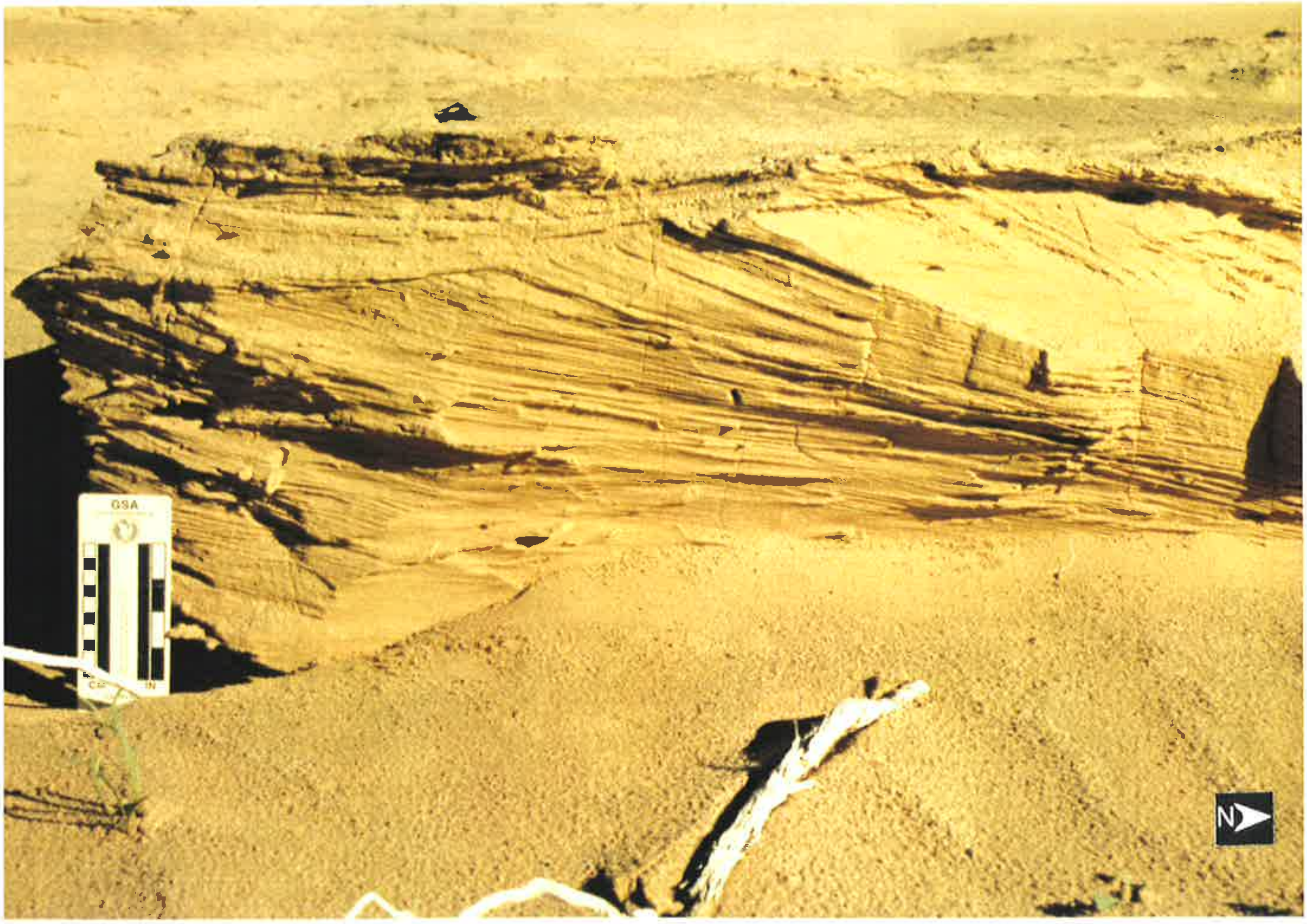

Figure 5.6 View of a flat-topped yardang showing an apparent 10°-20° dip of the relict linear dune's cross-strata. This may demonstrate lower dune-plinth accretion deposits. The flat clay top is a remnant of a previous desiccation polygon that influenced the genesis of the yardang form. A low angle (< 10° apparent dip) truncation or bounding surface is clearly shown in the upper 10 cm of the landform. An apparent northerly growth direction for the original linear dune is demonstrated by the preservation and exposure of the foreset beds.

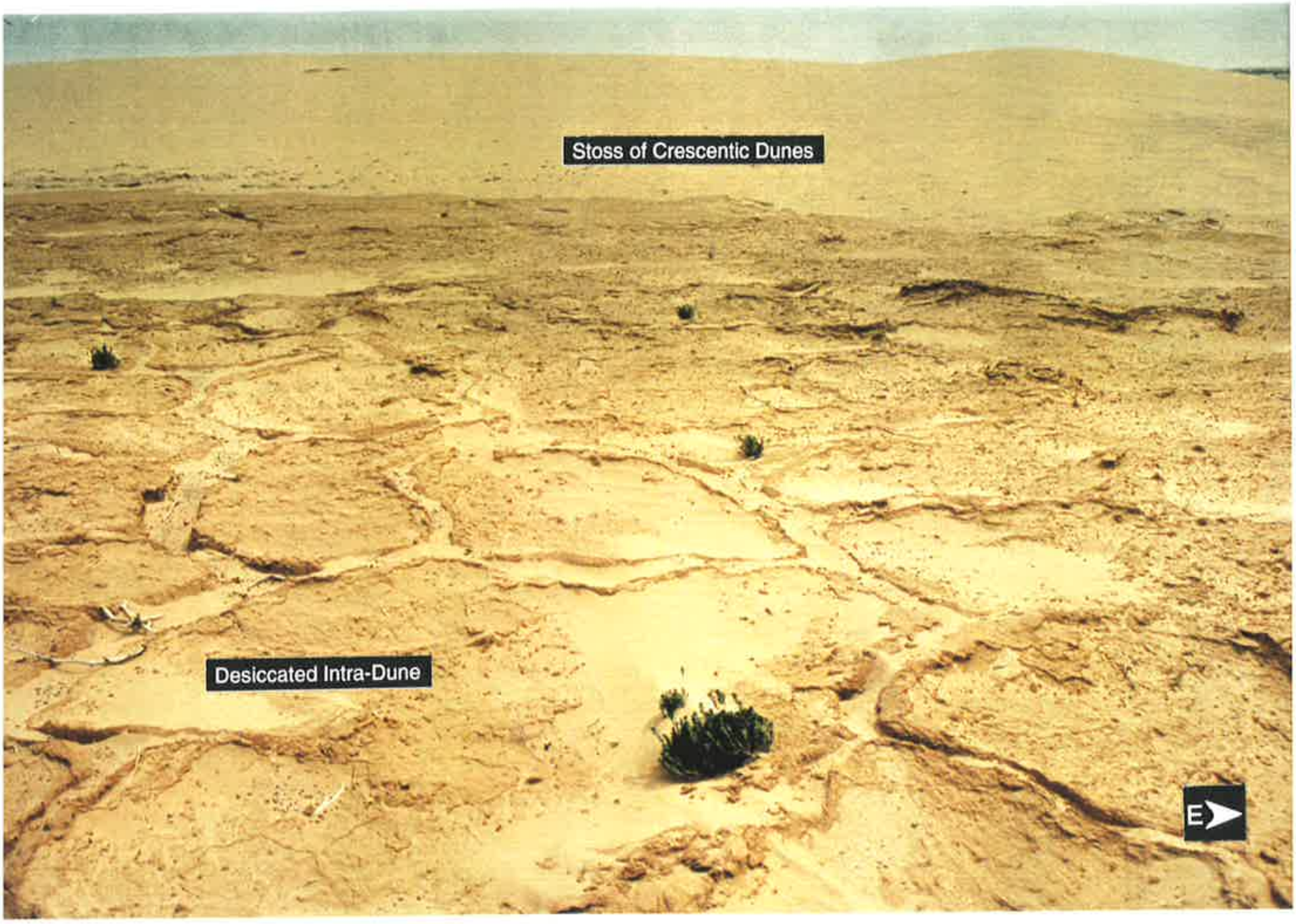

Figure 5.7 Non-orthogonal desiccation polygons (~0.5 m to 1 metre wide) at the southern end of the Gurra Gurra crescentic dunefield. In summer, the widening of the desiccation fractures occurs by intense southerly surface winds being channelled parallel to a polygon's perimeter. As the fracture widens, the infilling loose sand is transported downwind, which contemporaneously enhances downcutting and yardang shaping. Seasonal wet-dry and swell-shrink processes assist in causing the intradune to go through cycles of changing surface roughness over which sand patches can accumulate, air-flow is channelled, and sand is eroded and entrained into nearby evolving crescentic dunes. Combined with the significant dimensions of the linear dune, upon which upslope wind speed amplification can occur, the desiccation process is fundamental in both the genesis and supply of sand to the crescentic dunes.

The tapering form is more characteristic of yardangs that are of greater dimensions, and are indicative of up to 2 m of downcutting. Their upper surfaces characteristically taper leeward, with the planform showing an accentuated, sharp longitudinal crestline. This form is more typical of the shape of an inverted boat hull. The prow displays an erosional backcutting of the upper one-third to one-half of the feature, which often resembles a concavo-linear profile. As with the flat-topped form, the mid-section is wider and blocks of collapsed 'aeolinite' occur along the perimeter of taller features.

The sediments comprising these features are quartz-rich dune sands that have slight cohesive properties due to the presence of auxiliary clay cements. Examination by XRD has revealed that the cement and surface silt-clay 'skins' of the desiccation polygons comprise members from the 'swelling-clay' smectite group, specifically montmorillinite $(Na)_{0.7}(Al_{3.3}Mg_{0.7})Si_8O_{20}(OH)_4.nH_2O$, and kaolinite $Al_2O_3.2SiO_2.2H_2O$ of the kandite group of minerals. Sand cohesion, and hence yardang genesis, is assisted by the presence of these clay minerals, even though the weight percent of clay is < 1 % in any sample mass. Although yardangs are prevalent, due to their composition they are readily abraded and easily disaggregate under very low compressive forces, such as finger pressure. Due to the low sedimentary cohesion of these structures, the persistence of meso-yardangs under the multi-modal wind conditions at Gurra Gurra waterhole is *c.* ≤ 2 years, and represents a rapid rate of downcutting that occurs over sections of the linear dune upon which the crescentic dunes are building. The gentle nature of dune advance is demonstrated by their relationship with the yardang fields, where partially buried but preserved yardangs protrude from the windward slopes of some dunes. This attests to the gentle, non-erosive motion of a dune's advance, even under the energetic winds of summer.

5.5 Summary: Depositional and Erosional Landforms

Unequivocally, ripple forms and patterns, and the orientation of yardangs, nebkha and scour flutes have successfully determined the formative antecedent, seasonal wind conditions of the Gurra Gurra crescentic dunes. These results correlate with those discussed in Section 4.0, and identify a multi-modal aeolian regime affecting both the morphology and sedimentology of this localised dunefield.

Overall, all minor sedimentary landforms have demonstrated the existence of three prevalent wind directions of variable intensities. A robust primary southern wind regime occurs in summer and late spring, with a significant westerly component in the post-midday hours of the latter. Some evidence of this secondary vector is also found in summer. Contrary to this, the winter-early autumn regional wind pattern swings to a comparatively more gentle northerly air stream, in tandem with an intermittent and minor westerly component. It is this oblique westerly pattern that assists in the elongation of the eastern horn and the development of seif-linear form for some dunes. Similarly, the southern and northern flow regimes generate crescentic morphologies that cyclically pass between a state of equilibrium and quasi-equilibrium.

Clearly, slope gradient appears to direct the obliquity of ripple crestlines. Ripple morphology and orientation are consequences of slope angle, wind speed and direction, grain-size between intradune and crest, protection from air flow by the surrounding curvature of the barchanoid and the time of observation. Due to these many variables, ripple patterns upon inclined surfaces cannot be considered accurate current direction indicators. However, ripple patterns upon the sub-horizontal gradients of the crest, lower stoss and intradune are considered legitimate wind vanes.

Ripple indices were many and varied in all seasons (*RI* ~7-30). The dune ripples at Gurra Gurra waterhole do not have indices any higher than those of sub-aqueous origin, thereby, enforcing the principle that ripple indices are not diagnostic of environmental origin when bi- or multi-directional air flow occurs. This conclusion is also supported by Sharp (1963) for the Kelso dunes and Seppälä and Lindé (1978) from wind tunnel models.

It is apparent that desiccation of the intra-dune floor, combined with the cycle of yardang formation and destruction, are principle mechanisms whereby sand is sourced for the genesis and development of the crescentic dunes. Further discussion of the relationships between desiccation polygons and yardangs follows in Chapter 7.

6 Crescentic Dune Surficial Sedimentology

6.1 Descriptive Surficial Sedimentology

This chapter discusses how the granulometric distributions of the surficial sands comprising crescentic dunes are significantly controlled by changes in micro-geomorphic positions and, to a lesser extent, by the variable directions and intensities of seasonally-changing wind regimes. The fundamental sedimentological distributions are emplaced by the primary summer wind regime. Lesser influential aeolian trends cause only minor alterations to the principal sedimentological distributions, with the surficial sedimentology of the crestal area being most altered by upslope wind speed amplification. However, only by finely-tuned sampling and comparative analytical techniques can such subtle changes be unveiled.

The mineralogy of the Gurra Gurra dune sands is predominately quartz ($\geq$ 95%) with accessory components of ferruginous lithic fragments or 'micro-conglomerates', tourmaline and plagioclase, with very minor zircon and rutile occurring in only the fine to very fine fractions (2ϕ - 4ϕ). The accessory fines accumulate as deposits that delineate ripple patterns into coarse, quartz-rich crests and darker colored heavy mineral-rich troughs. Figures 6.1-6.3 are representative histograms of sediment characteristics for each different geomorphic setting of the study site in each season. Distributions within the medium-to-fine grain-size range, from both linear and crescentic dunes, exhibit unimodality, positive skewness and leptokurtosis. This sedimentological character typifies all seasons and is indicative of the crescentic dunes being built primarily from sands derived from the underlying linear dune. This phenomenon characterises a predominantly semi-closed sedimentologic system, as very minor mass transfer enters or leaves this small dune-field. Although Strzelecki Creek offers an abundant sediment supply for dune building, there is apparently minimal volumetric input into the crescentic dune-field. The sediment of the dunes is characterised by an iron oxide coating (patina), whereas, the river sands typically show < 10 per cent of grains to be covered by patinas. The young age of only decades for

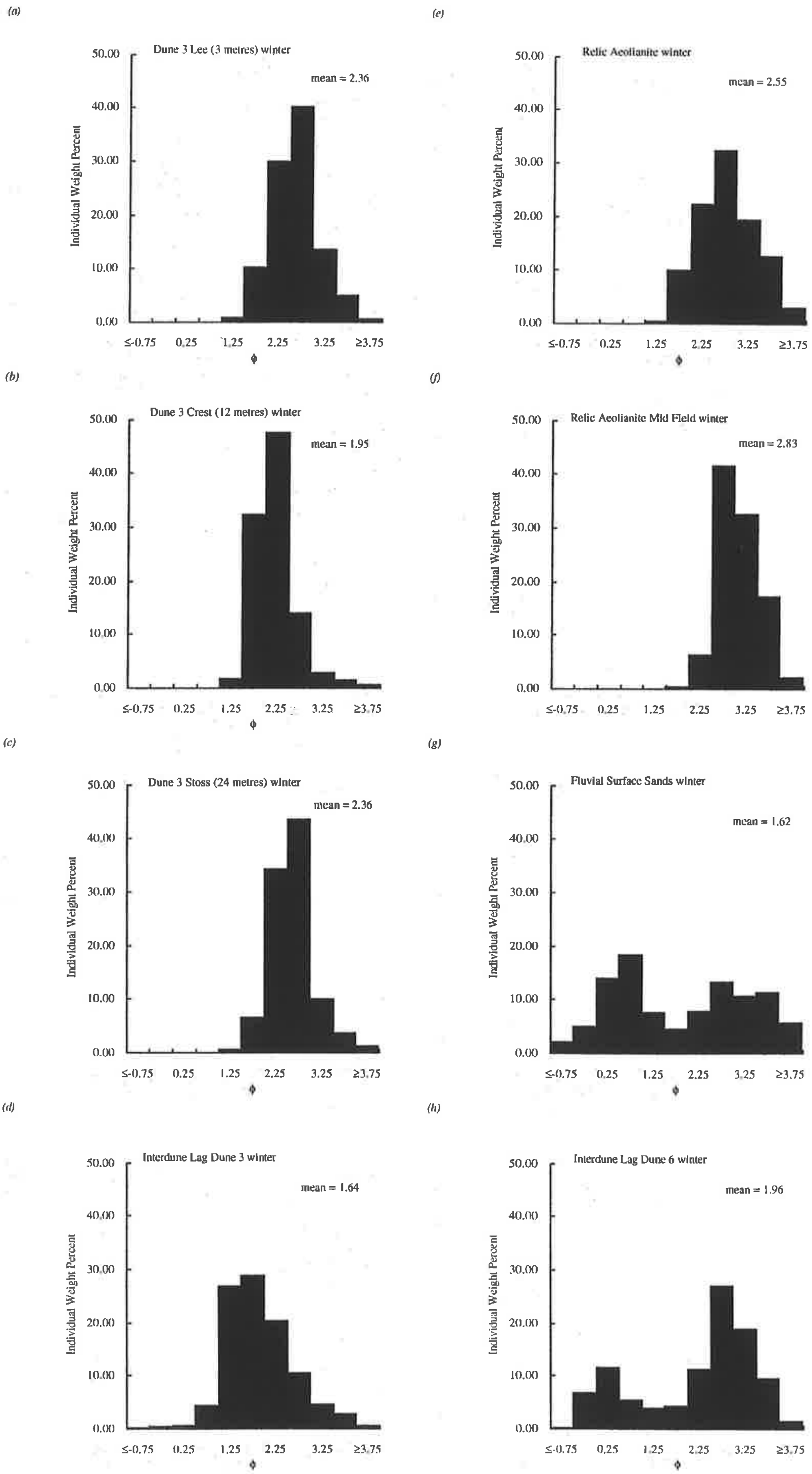

Figure 6.1 Representative histograms of the winter crescentic, linear and fluvial micro-environments. Note the bimodality of both the fluvial and interdune deposits, as well as the typical unimodal nature of both the linear and crescentic dune sands.

the crescentic dune building episode (see Section 4.1.1) may provide insufficient time for iron oxide coatings to have developed upon fluvially-derived grains. Evidence against the crescentic dune sand being derived from modern-day fluvial sediments is provided by:

- a distinct difference in sand grain Munsell colour for the same mean grain-size range from the separate sedimentary domains.
  - fluvial colour ~10YR 7/3 is very pale brown
  - crescentic dune colour ~7.5 YR 6/6 is reddish-yellow
  - linear dune 'aeolinite' colour ~7.5YR 6/6 is reddish-yellow
  - interdune 'reg' ~7.5YR 7/4 is pink
- the dominance and annual prevalence of sand drift potential by southerly winds which are out of sympathy with the required sand transport direction for a fluvial sediment source.
- relict linear dune lowering and form mutilation is apparent with the genesis of the crescentic dunes.

It is most probable that alluvium of an antecedent Strzelecki Creek and associated flood plain acted as sources for the relict linear dunes of this region some 14,000-to-25,000 year B.P. (Wasson, 1983a). In winter, only the corridors of the transverse dunes bordering the creek (*c.* 200m distant) show evidence of fluvial sand input (Fig. 6.1h). Such input is characterised by coarse reg-like deposits emplaced by intermittent northerly storm 'bursts' and associated increased saltation. In spring, a slightly increased input of fluvial sands also occurs, with the existence of coarse, translucent-grain deposits within the interdune corridors and the perimeters of the crescentic dunes (Fig. 6.3a-b). Significantly, these sands are found throughout the entire field of crescentic dunes (> 2 km from the source) and represent a higher weight percentage of coarser sands in the surficial interdune and intradune sediments than under winter conditions (Fig. 6.3g-h and 6.1h). This reflects the dominance of the wide drift potential of westerly spring winds (Fig. 1.3 of Fryberger, 1979), and the location of the dune field almost within a meander loop of Strzelecki Creek (see earlier Fig. 1.4, Chapter 1), from which some alluvial sand enters the crescentic dune-field.

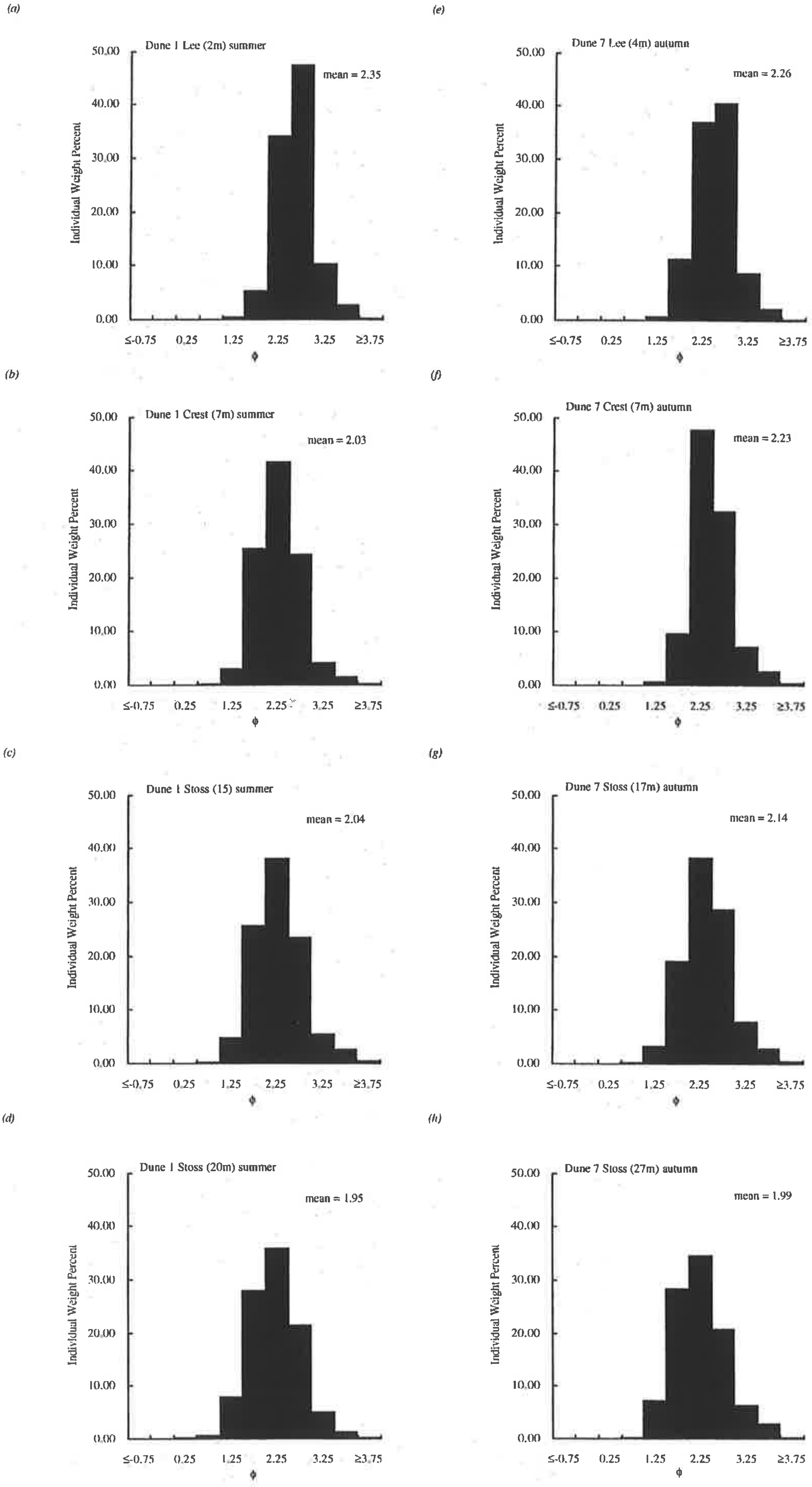

Figure 6.2 Representative histograms of summer and autumn dune sediments.

It could be argued that fine, medium and coarse sands are derived from the alluvium and are intermixed with the linear materials. Possibly this does occur to some extent, but at this stage of development of the crescentic dunes, there has been little effect on the colour range of the sands. The sands of the linear dune and crescentic dunes maintain identical Munsell colours (7.5YR 6/6), while the dominant annual aeolian transport direction is significantly north and therefore toward rather than away from the fluvial source. Furthermore, the fine and medium sand fractions are prevalent below the coarser sand grades of the surface laminae for the intra-channel alluvium deposits and are not readily available for saltation. Most likely, inter-grain sorting of fines occurs between the larger grains which protects and concentrates or grades the finer materials below the coarser surface laminae. Only under extremely high threshold velocities (a rare event from the northerly directions) do both coarse and fine alluvium saltate into the crescentic dune area. For these reasons, it is most unlikely that the crescentic dunes are derived chiefly from the modern day alluvium.

Also notable is the greater abundance of coarser fractions in the surficial fluvial sands of spring when compared to winter (Fig. 6.1g and 6.3e). This may indicate that in spring, preferential vertical grading (sifting and size segregation) under strong winds induce vibrations, caused by inter-particle motion and collision, which develops a finer assemblage of sands at depth beneath a surficial layer of dominantly coarse materials. Under such conditions, the coarser grades are preferentially transported to the interdune area (Fig. 6.3e-h). In winter under predominantly quiescent conditions, a similar mechanism is involved. Finer sands are protected between the intergranular voids of the larger grains, but are not concentrated at depth due to reduced impact energies and impact frequencies. Increased quantities of finer materials are therefore available for transport into the crescentic dunefield with increased shear velocities. This may account for the greater weight percent of fines found in the winter distributions.

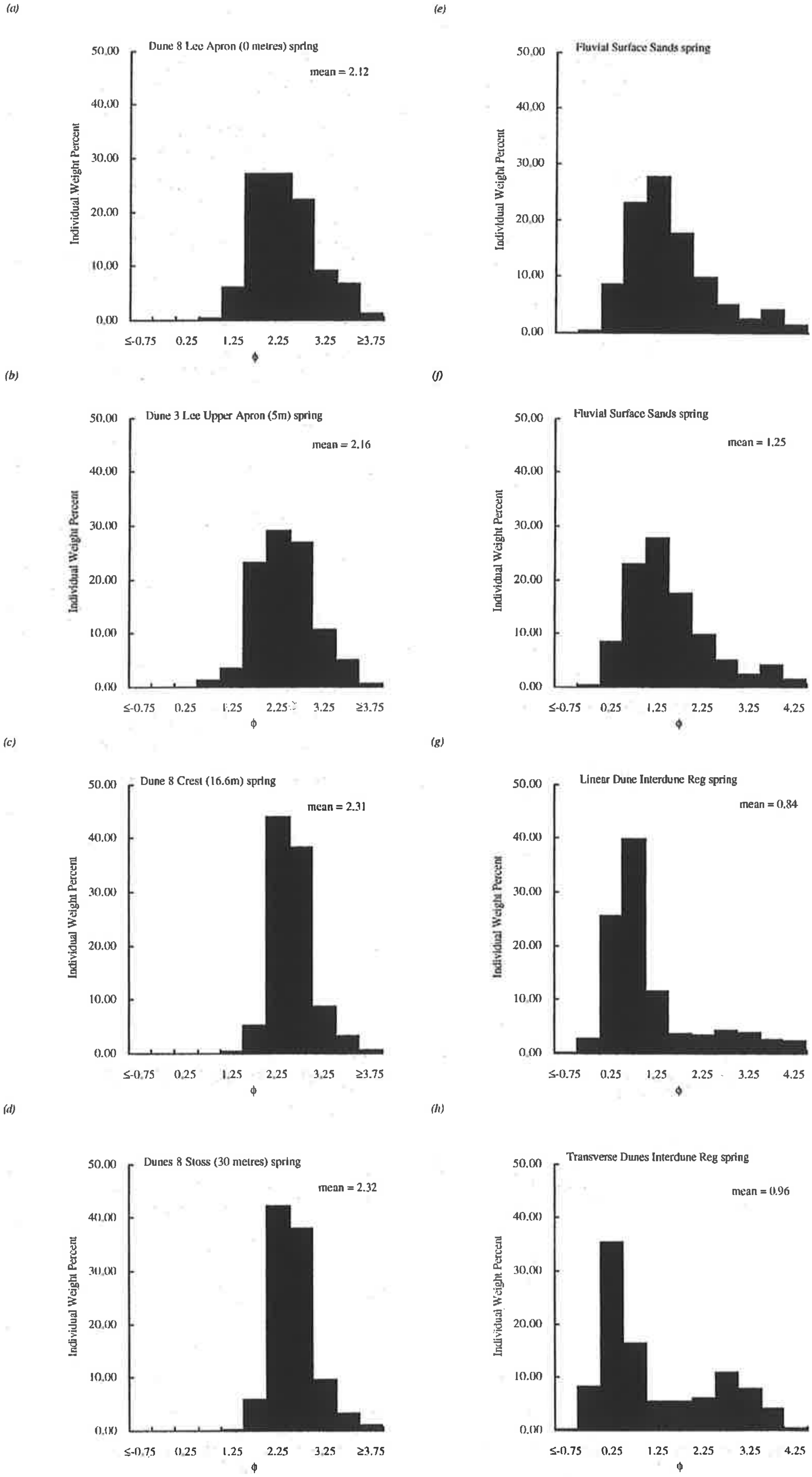

Figure 6.3 Representative histograms of spring dune sediments. Note the bimodality and relative coarseness of the fluvial and interdune sands

No distinct differences in mean grain-size, sorting, skewness or kurtosis are readily shown via sand histograms and the general descriptive statistics of samples from the crescentic dunes. This is especially the case when multiple dunes are examined in two or more seasons. In such instances the interrelationships and influences of different dunes and different seasons are not easily distinguished. It is, therefore, appropriate to examine the relationships between spatial (morphologic location) and temporal (seasonality) frames of reference using comparative statistical methods, which test the significance and validity of any apparent differences. The simplest method from which spatial and temporal changes can be observed, is by the examination of moment measures as graphed in Fig. 6.4 - 6.31, Appendix 3. It is seen that explicit differences of sedimentology occur with morphologic location, as well as changing season. These differences, their magnitudes, validity and significance will now be explored.

6.2 Qualitative Spatial Contrasts of Surficial Sedimentology

6.2.1 Winter

Sedimentological distributions for the Gurra Gurra crescentic dune sands lie solely within the medium-to-fine grain-size range, are unimodal, positive skewed and leptokurtic (Fig. 6.4 - 6.31, Appendix 3). Mean grain-size for all samples lies within the fine to medium grade classes, with a predominance of fine grades. The crest-brink zone (crest, upper lee and upper stoss positions) are coarsest [Md = 1.97ϕ (0.26 mm)] relative to other positions on the dune (Fig. 6.4), while the basal stoss shows the finest grain-sizes [Md = 2.33ϕ (0.205 mm)] for the majority of samples (dunes 1, 3, 4, 5). Exception to this is shown by dunes 2 and 6 where relatively coarser toes are evident. However, a fine lee base [Md = 2.24ϕ (0.21 mm)] is characteristic of all dunes. All dunes display a coarse to fine downslope grading on the lee, with an overall higher degree of coarseness compared to the stoss. A broad scale gradation of fine to coarse occurs upslope along the stoss for dunes 1, 3, 4 and 5, while dunes 3 and 6 again differ, with irregular grain-size variation upslope.

The horns of the winter dunes are composed of fine to medium grade sands where the range of mean grain-size is 2.71ϕ (0.15 mm) to 1.62ϕ (0.33 mm) (Fig. 6.20), with a median of 2.15ϕ (0.23 mm) and 2.29ϕ (0.205 mm) for the western and eastern horns respectively. Although a considerable degree of variation is found on both horns, as well as for individual dunes, a general trend of coarsening from the finer basal locale towards the coarser higher elevations is apparent for all dunes.

Moment:Location	Winter	Summer	Autumn	Spring
1 lee	2.24	2.35	2.26	2.34
1 crest	1.97	2.15	2.23	2.12
1 stoss	2.33	2.09	2.03	2.16
1 horn west	2.15	2.07	2.36	2.19
1 horn east	2.29	2.10	2.04	2.19
2 lee	0.50	0.43	0.45	0.50
2 crest	0.47	0.52	0.45	0.45
2 stoss	0.47	0.57	0.54	0.52
2 horn west	0.47	0.50	0.46	0.50
2 horn east	0.45	0.49	0.55	0.46
3 lee	0.45	0.32	0.21	0.42
3 crest	0.83	0.68	0.67	0.74
3 stoss	0.47	0.35	0.44	0.43
3 horn west	0.57	0.52	0.34	0.48
3 horn east	0.63	0.57	0.73	0.62
4 lee	3.57	3.70	3.35	3.49
4 crest	4.62	4.29	4.12	4.21
4 stoss	3.79	3.67	3.39	3.57
4 horn west	3.81	3.75	3.50	3.53
4 horn east	3.93	3.92	4.09	3.68

Table 6.1 Seasonal median (ϕ) values of granulometric moments for different micro-morphologic elements.

Sorting for the winter dunes lies within the grades of moderately well sorted and well sorted. Sorting is gradational along the lee slope (Fig. 6.5) and is best on the upper lee position for dunes 1, 4, 5 and 6 while dunes 2 and 3 show the upper stoss position to be the better sorted locality. As with mean grain-size, sands having greater degrees of sorting occur on the highest positions of the dunes. The lee slope is characterised by gradational and poorer sorting downslope, while the stoss element shows random distributions of sorting along its length. Generally the degree of sorting is low at the base of both the stoss and lee slopes nearest the intra-dune area for dunes 1, 2, 3, 5 and 6, with the worst sorting being at the approximate mid-stoss and rear slipface (stoss) positions for dunes 1, 3 and 5. However, very little difference is found between the median values of sorting for any micro-morphologic locations (Fig. 6.5 and Table 6.1).

Dispersion along the horns is shown by well to moderately well sorted sands, with a range between 0.63ϕ to 0.37ϕ. As with grain-size, there are no apparent dispersion differences for either horn, where a high degree of variability offers no distinct trend of change. A hint of better crest sorting for dunes 1, 3 and 4 may be observed. In general, however, sorting is random over the dune horns, with neither crestal nor basal elements being any better sorted than each other (Fig. 6.21). Skewness of the winter sediment samples show an excess of fine materials (i.e. fine or positive skew, Fig. 6.6). The upper morphologic localities (crest, upper lee and upper stoss) of the separated brink and crest demonstrate the regions of greatest fine skew for all dunes (1 - 6) (Md = 0.83). The lee slipface also portrays a gradational decrease of skewness downslope for dunes 1, 2, 4, 5 and 6 (Md = 0.47). The stoss slope is rather variable (Md = 0.47), but with a general trend of increasing skewness upslope.

Skewness for the horns is predominantly positive (fine), with the crestal regions generally having the highest degree of skewness. Skewness values range between 1.09 and 0.04; the low morphologic elements are least skewed compared to the higher degree of skewness at the crest. Most sample positions are very finely skewed with values > 0.50.

No apparent differences are recognised for the eastern or western horns (Fig. 6.22). Kurtosis values for all samples are typically leptokurtic or peaked. Exceptions are found with the platykurtic nature of the lower positions of the lee and stoss of dunes 2 and 6. Greatest peakedness of the distributions is demonstrated at the upper dune positions near and inclusive of the crest (Md = 4.62). The lee slope demonstrates a grading of leptokurtosis from highest at the top of the slope to lowest at the base (Md = 3.57). The stoss slope shows a general high degree of variation along its length, with a distinct decrease in leptokurtosity for the samples taken within the domain of the stoss rear slipface (Md = 3.79). The kurtosis of the winter horns shows a vast range (8.10 highly leptokurtic to 2.95 slightly platykurtic) that can be determined using moment calculations. All samples except one for each of dunes 3 and 4, are varyingly leptokurtic, with the crestal region showing a consistently high kurtosis. Although no significant differences are readily apparent between the eastern and western horns, incremental changes across each horn that culminate in a higher crestal kurtosis, apply for dunes 3, 4 and 5.

6.2.2 Summer

The sediment sample distributions of summer reflect distinctive profile differences from those of winter (Fig. 6.8). All sediments lie within the Wentworth grades of fine to medium, with the lee slope of all dunes 1, 2, 3a, 3b and 6 composed of fine sediments (Md = 2.35ϕ, 0.20 mm). The crestal zone is the locality of mean grain-size intermediate (Md = 2.15ϕ, 0.23 mm) between the finest sands of the lee-slipface and the coarser composition of the stoss element (Md = 2.09ϕ, 0.235 mm). Comparatively, the stoss slope shows granulometric variation between both grades. A definite and sharply accentuated fining of the lower stoss regions for dunes 2, 3b and 6 is clearly expressed, where grain-size is similar to that of the crest. The mean grain-size distributions for the horns of the summer dunes are similar to those of winter, with both being composed of fine to medium grain-sizes ranging between 2.70ϕ (0.15 mm) and 1.62ϕ (0.32 mm). Similarly, the incremental change of grain-size from the finer, lower elements to coarser crests of dunes 1, 3a and 6, is a characteristic of both seasons (Fig. 6.24). The western horn also shows a generally coarser nature than the eastern horn on dunes 1, 2, 3a and 3b.

Sorting, as demonstrated in Fig. 6.9, lies between the grades of moderately well sorted and well sorted. The crestal positions, as with mean grain-sizes, are generally intermediate (Md = 0.52ϕ) between the better sorted lee (Md = 0.43ϕ) and least sorted stoss (Md = 0.57ϕ) slopes. Overall, an incremental trend from moderately well sorted towards well sorted occurs away from the stoss toe. Although some variation is found upon the stoss slope, this deviation is not significantly different from the overall trend of better sorting over the dune from stoss to lee. Sediments (Fig. 6.25) along a leeward transverse profile are moderately well to well sorted with a high degree of variability (range 0.64ϕ to 0.41ϕ) for both horns, without any common trend. For example, dune 3a depicts better sorting from east to west, dune 6 has better sorting on the crest relative to the lower horns, while dune 3b displays no change on the western horn between interdune and crest areas, but shows a distinctive linear increase of sorting from crest to interdune for the eastern horn.

Skewness of the summer sediment samples on dunes 1, 2, 3b and 6 is positive (fine), with an exception occurring with one stoss sample of dune 3a. The basal regions of both the lee and stoss elements are least skewed (Md = 0.32 and 0.35, respectively), while the upper micro-morphologic positions show variable but a high degree of fine skewness (Md = 0.68, Fig. 6.10). Excepting for dune 3a, where a coarsely skewed sample is recorded, there is a consistent and distinctive increase of fine particles and hence positive skewness upslope on the stoss, and a trend of decreasing skewness downslope on the lee. In common with the winter developed horns, those of the summer are prevalently finely skewed (Fig. 6.26). The range of skewness is also consistent with winter values, lying between 0.99 and 0.04, again with most samples being above a value of 0.50. Due to the considerable variability between the values of skewness on both eastern and western horn, no discernible trends are obvious.

Kurtosis values for all samples are profoundly leptokurtic (Fig. 6.11), except for one basal stoss sample from dune 2, which is normally distributed. The crestal region displays the highest kurtosis values (Md = 4.29) on dunes 1, 2, 3b and 6, while low leptokurtosis values are distinguished on the lower stoss of dunes 1, 2, 3a, 3b and 6 (Md = 3.67)., and in several instances, the lower lee portions of dunes 2, 3a and 3b (Md = 3.70). Transverse to the crestline, kurtosis is predominantly leptokurtic and as with skew and sorting, displays a high degree of intra-element variability (Fig. 6.27).

6.2.3 Autumn

Not unlike the previous results of winter and summer, mean grain-sizes for the autumn samples lie within the Wentworth grades of fine and medium sand. Noticeable is the identification of relative crestal fining (Fig. 6.12) for dunes 3, 7 and 8 [Md = 2.23ϕ, (0.22 mm)]. Upper micro-morphologies consist typically of fine grades, as does the lee slope [Md = 2.26 ϕ (0.21 mm)] which shows an apparent coarsening downslope from the crest. The lee does not, however, greatly deviate from the mean grain-size values of the crest, whereas greater variation is shown on the stoss element [Md = 2.03ϕ (0.25 mm)]. In general, a gradual but highly variable trend of decreasing grain-size is observed from the stoss toe towards the crestal position. Significant is the presence of a fine grade sand at the dune toe on dunes 3, 7 and 8.

Fine to medium sand fractions are prevalent between the range of 2.65ϕ (0.16 mm) to 1.98ϕ (0.25 mm), which characteristically increase in mean grain-size from the western horn base to the eastern horn base. The finest sands are found upon the lower western horn, grading up and over the intermediate mean grain-size of the crest to the coarsest sands on the lower eastern horn. All micro-morphologic positions of the western horns are finer than for any locality on the eastern horns. This may represent the occurrence of oblique transport between the forward and reversed longitudinal flow regimes of summer and winter, respectively.

These samples are moderately well to well sorted (Fig. 6.13). The poorest sorting is found on the stoss slope (Md = 0.54ϕ) where the base or toe of the stoss is least sorted, with gradually improved sorting occurring upslope over the crest and down the slipface (Md = 0.45ϕ) for both morphologic elements. The base of the lee shows poorer sorting as compared with the mid and upper lee positions, but is still better sorted than the entire stoss slope. The autumn horn samples are moderately well to well sorted (range 0.55ϕ to 0.36ϕ) and follow a definite trend of poorer sorting on the eastern horn relative to better sorting on the western horn (Fig. 6.13). The crest is intermediate between the two extremes. An association exists between mean grain-size and sorting; mean grain-size progressively increases with decreased sorting. This characteristic is only obvious with autumnal sands, as both winter and summer horn sediments depict random distributions overall.

Skewness for autumn is positive (Md = 0.67) for all dune samples regardless of micro-geographic location, although the highest degree of skew occurs on the crest for all dunes. The lee slope is least finely skewed (Md = 0.44), while the stoss gradient is variable (Md = 0.44) and intermediate between the crest and lee, with a general decrease downslope towards the intra-dune area, for both morphologic elements (Fig. 6.14). Skewness for the autumn horns is finely to very finely skewed (1.00 to 0.30) and, as for trends found with both mean grain-size and sorting, skewness is seen to be greater on the eastern horn compared to the less skewed western horn. The samples closest to the interdune areas for both horns also demonstrate the lowest quantity of skew for each horn.

Kurtosis is typically leptokurtic (Md = 4.12) with the highest values found on the upper morphologic regions (Fig. 6.15). The base of the stoss slope is least peaked with all dunes approaching zero values. The toe of dune 3 actually becomes platykurtic. The base of the lee slope similarly shows less kurtosis than the upper micro-geomorphologic positions. The trend for kurtosis is for increasing leptokurtosis upslope of the stoss toe (Md = 3.39) and decreasing leptokurtosis downslope of the slipface (Md = 3.35). Leptokurtosis is also characteristic of upper and lower micro-morphologic locations on the horn and mirrors the profile of horn skewness, the lowest values being associated with the lower horns.

Of all the seasonal transverse profiles, those for autumn show the most distinctive differences (Fig. 6.12 - 6.15). However, caution must be exercised in the interpretation of this data, due to the low sample number.

6.2.4 Spring

The mean grain-size for spring (Fig. 6.16) shows a predominance of fine grade sands for dunes 2, 7, 8, 9 and 10. Median grain-size ranges from 2.34ϕ (0.20 mm) on the lee, 2.12ϕ (0.235 mm) on the crest and 2.16ϕ (0.23 mm) on the stoss. However, the mid-to-upper lee shows the finest compositions relative to the crest and stoss. Grain-size variation is significant at the base of the lee and toe of the stoss for individual dunes. For example, dunes 2 and 9 show the base of the lee to comprise the finest sands. However, dunes 7 and 8 show this position to be composed of the coarsest grade sands (Fig. 6.16). Similarly, the toes of dunes 2, 8 and 9 are coarse compared to the fine sands of dunes 7 and 8. The stoss again shows the characteristic signature of variable fluctuating grain-sizes, although these are more strident in spring than in other seasons. Spring, like all previous seasons, demonstrates the Wentworth mean grain-size range (2.55ϕ, 0.17 mm to 1.93ϕ,0.26 mm) to lie in the fine and medium fractions (Fig. 6.28) along the horns. Indefinite indication of coarsening towards the crest is hampered by the high degree of variability on each of the horns. No apparent distinction is possible between the east and west horns as too few dunes are available for comparisons to be effective.

Spring samples are moderately well sorted to well sorted (Fig. 6.17), with the upper morphologic localities being better sorted (Md = 0.45ϕ) than the basal parts of the dunes. Sorting along the stoss elements is poorest at the toe and improves upslope for all dunes (Md = 0.52ϕ). Slipface sorting is not the same for all dunes, with dunes 7, 8 and 10 showing decreased sorting downslope with a distinct near-linear trend, while dunes 2 and 9 show a fluctuating deviation (Md = 0.50). Horn samples are moderately well to well sorted, with results showing better sorting on the higher positions (Fig. 6.29).

Skewness is shown to be positive for all samples (Md = 0.74), with the highest degree of fine skew found on the upper micro-morphologic positions (Fig. 6.18). A high degree of variability exists for the third moment with each of the dunes, making it difficult to determine whether a particular morphologic element has a specific degree of skewness. Scrutiny by statistical methods is required to quantify whether the stoss samples (Md = 0.43) are more or less finely skewed than those from the lee (Md = 0.42). The higher micro-geomorphic locations are more finely skewed than the lower horns (Fig. 6.30).

The crestal regions of the dunes display an extremely leptokurtic nature (Md = 4.21), (Fig. 6.19). The toe of the stoss consistently shows low values of kurtosis for dunes 2, 7, 8, 9 and 10, with dunes 7, 9 and 10 entering the platykurtic field (Md = 3.57). Likewise, the lee slopes of dunes 7, 8 and are partially within the platykurtic field (Md = 3.49). Individual dune variability is again large, with only the crestal locality significantly different when using graphical (qualitative) analysis. Leptokurtosis is also prevalent on the horns and reflects a pattern similar to skew. That is, the crestal zone is one of high leptokurtosis, whilst lower values are found on the lower regions.

6.3 Quantitative Spatial Variation

To further define, enhance and test the significance of sedimentological change as identified by qualitative analysis between the different micro-morphological locations in each season, a subsequent examination was conducted using two versatile non-parametric tests, the Kruskal-Wallis and Scheffé Projection techniques. Qualitative analysis clearly discerns that differences occur between the upper and lower micro-morphologic localities, as shown by both the graphical trends and the degree of differentiation between median values (Table 6.1). The Kruskal-Wallis H statistic is used to test the significance of these differences for like moments between different morphological elements (low lee, high crest-brink, low stoss, low horn-west, low horn-east) for individual seasons.

Table 6.2 illustrates the *H* statistic and the level of significant difference between granulometric parameters and k=5 group morphologic elements for individual seasons. The *H* statistic for mean grain-size between the five morphologic elements exceeds the critical value for winter, summer and autumn, but not spring samples. The larger the statistic the less likely the elements are to have the same or similar mean grain-sizes. This distinction is most obvious for autumn samples where the statistic is 15.62, less so for winter and summer (H = 12.17 and 12.02 respectively) and least for spring (H = 7.35), when compared with the critical value of 9.49 at the 0.05 level of significance.

Season	Moment	H statistic	Significance Level	Scheffé Projection (k-groups)
winter	1	12.17	0.05	(1&2)
winter	2	12.01	0.05	(1&3) (3&5)
winter	3	24.57	0.01	(1&2) (1&3)
summer	1	12.02	0.05	(2&3)
summer	2	21.20	0.01	(2&3) (2&5)
summer	3	14.75	0.01	(2&5) (3&5)
autumn	1	15.62	0.01	(1&2) (2&3) (3&5)
autumn	2	13.29	0.01	(2&3) (3&5)
autumn	3	20.35	0.01	(1&2) (1&3) (2&5) (3&5)
spring	1	7.35	not significant	no differences
spring	2	23.01	0.01	(1&2) (1&3) (1&4)
spring	3	19.03	0.01	(1&2) (1&3)

Table 6.2 Seasonal interpretation of significant differences between different morphologic elements, where: df=4 and k=(crest-brink = 1, low stoss = 2, low lee = 3, low horn west = 4 and low horn east = 5). For a significant difference to occur and to be tested using the Scheffé projection method, H ≥ critical value. Critical values are $\chi^2_{0.05} = 9.49$ and $\chi^2_{0.01} = 13.28$. The larger the test statistic the greater the difference between two or more elements.

In all cases, except for spring, the null hypothesis of no difference between the micro-geomorphologic elements of (low lee, crest-brink, low stoss, low horn-east, low horn-west) is rejected. The H statistic exceeds the difference that is likely to have occurred by chance. For spring, no mean grain-size difference is found between the micro-morphologic elements.

No precise nature of the differences is given by the Kruskal-Wallis test, however, it is apparent that significant differences in mean grain-size do occur at different morphological localities over a crescentic dune. It cannot be ascertained whether all morphologic elements are different from one another, or whether two or more elements are drastically different from the others. To deduce a solution to this problem, a Scheffé type projection was performed, whereby individual *k*-group differences could be identified. Obviously, if no difference is expounded by the Kruskal-Wallis test, then the application of the Scheffé type projection is unwarranted. Table 6.2 demonstrates the results of the Scheffé projection between *k*-groups.

It is apparent that mean grain-size[1] is significantly different between the morphologic elements:

- crest-brink (coarse) and low stoss (fine) in winter
- low lee (fine) and low stoss (coarse) in summer
- crest-brink (fine) and low stoss (coarse) in autumn
- low stoss (coarse) and low lee (fine) in autumn
- low lee (fine) and low horn-east (coarse) in autumn

All other combinations of *k*-groups show no significant differences from the calculated critical values at the 0.05 level of significance and, therefore, are not dissimilar in mean grain-size. The null hypothesis of no difference cannot be rejected in these cases. For example, the winter crest-brink is shown to be the coarsest morphologic element and the

[1] Note that the terms coarse - fine, better - poorer, less positive and more positive are relative expressions of values, and are not absolute values based upon any grade or scale system.

low stoss the finest (1 and 2). There is no difference between any other combinations of

the comparative element sets. i.e. $\begin{matrix} (1and3) & (2and3) & & \\ (1and4) & (2and4) & (3and4) & \\ (1and5) & (2and5) & (3and5) & (4and5) \end{matrix}$

The general lack of significant differences between the majority of element combinations reveals the crest-brink and the low stoss elements to be end-members of a very narrow mean grain-size range and that the dune sand is very refined and limited in granulometric variability. Subsequently, the relative range of the moments can be expressed as a series where each end member is the first (fine) and last (coarse) element, with a sequence of incremental variability between them. For example, variability of relative fine to coarse mean grain-size can be expressed as:

- winter = (stoss, lee/he/hw, crest)
- summer = (lee, crest/he/hw, stoss)
- autumn = (lee, crest, hw, he, stoss)
- spring = no differences

These expressions illustrate that for winter, the low stoss element is finer than the lee, both horns and crest, while no significant differences exist between the lee, horns and stoss, or the lee, horns and crest. A similar series occurs for summer. However, when a greater number of element pairs show variability, as in autumn, a more sequential grading of relative granulometric character is possible. Here a series of incremental changes can be established by the inter-comparisons of morphologic elements. Again the relative end-members are fine and coarse, but in this case, the statistical projections determine how close each element is to another. Immediately obvious is the greater variability of the autumn results than those for either winter or summer. Note, that the separation of all elements into components separated by a comma, does not represent a significant difference between all separated elements within the set, but rather the most probable series of variation for those elements, some of which are significantly different from each other, the further they are apart. This procedure simply allows quick referencing between the upper and lower limits of a particular moment measure, and is a synopsis of both the graphical and statistical modes of analysis.

Sorting is also shown to be significantly different for some micro-geomorphologic elements in each season (Table 6.1). The *H* statistic demonstrates that the null hypothesis can be rejected, as $H \geq$ critical value. The greatest distinction of difference is demonstrated by the test statistic of summer (21.20), thereby indicating the greatest degree of difference between two or more morphologic elements.

Table 6.2 shows the results of the Scheffé projection of the Kruskal-Wallis test and demonstrates that sorting is significantly different between the morphologic elements:

- crest-brink (better) and low lee (poorer) in winter
- low lee (poorer)and low horn-east (better) in winter
- low stoss (poorer) and low lee (better) in summer
- low stoss (poorer) and low horn-east (better) in summer
- low stoss (poorer) and low lee (better) in autumn
- low lee (better) and low horn-east (poorer) in autumn
- crest-brink (better) and low stoss (poorer) in spring
- crest-brink (better) and low lee (poorer) in spring
- crest-brink (better) and low horn west (poorer) in spring

The series of relative sorting for each element from better to poorer for each season is:

- winter = (crest, horn east, stoss/horn west lee)
- summer = (lee/horn east, crest/horn west, stoss)
- autumn = (lee, crest/horn west, horn east, stoss)
- spring = (crest, horn east, lee/horn west, stoss)

Skewness between inter-groups is demonstrated to be significantly different at the 0.05 level for all seasons. Variation in the third moment and micro-geomorphic location occur between the:

- crest-brink (more +) and low stoss (less +) in winter
- crest-brink (more +) and low lee (less +) in winter

- low stoss (less +) and low horn-east (more +) in summer
- low lee (less +) and low horn-east (more +) in summer
- crest-brink (more +) and low stoss (less +) in autumn
- crest-brink (more +) and low lee (less +) in autumn
- low stoss (less +) and low horn-east (more +) in autumn
- low lee (less +) and low horn-east (more +) in autumn
- crest-brink(more +) and low stoss (less +) in spring
- crest-brink (more +) and low lee (less +) in spring

The series of skewness from more positive to less positive for each season is as follows:

- winter = (crest, he/hw, stoss/lee)
- summer = (he, crest/hw, stoss/lee)
- autumn = (crest/he, hw, stoss/lee)
- spring = (crest, he/hw, lee/stoss)

The Scheffé projection has demonstrated that in many instances, the null hypothesis of 'no difference' must be rejected, and that mean grain-size, sorting and skewness vary with micro-geomorphologic position on a crescentic dune.

6.4 Summary of Spatial Sedimentological Change

Qualitative interpretation of sedimentological profiles, (Appendix 3), coupled with statistical tests, have identified the significance of granulometric differentiation, respective of micro-geomorphic position, for the surface sedimentology of the Gurra Gurra crescentic dunes. The sedimentological signature of winter implies a correlation of granulometric parameters and micro-geomorphologic location. That is:

- mean grain-size is greatest and sorting is best within the crestal domain of the dune compared to the lower lee, lower horn and lower stoss positions;

- the degree of fine skewness and the peakedness of the sample distributions are comparatively greater at the crest-brink.

Similarly, the associations for the sedimentology of summer are:

- the mean grain-size of the stoss slope is relatively coarse compared to both the intermediate coarseness of the crest and horns, and the finer lee slope;
- a strong suggestion of dependency between better sorting with decreasing (finer) grain-size;
- both skewness and kurtosis are least at the lower stoss and lower lee position; the upper dune and lower western horn are comparatively higher with the lower eastern horn being highest;
- maximum skewness is not correlated with either the finest or coarsest mean grain sizes, but is intermediate between them; this is a contrast to both the first and second moments where change is linear.

The distributions of autumn are summarised as showing:

- coarser sands are prevalent on the lower stoss and horns, while the crest-brink is composed of relatively finer sediments;
- better sorting is associated with the finer grain-sizes, and show an inversely proportional relationship with the variation in mean grain-size; i.e. as mean grain-size increases, the degree of sorting decreases;
- extremely positive skewness is associated with the finer sands of the crest-brink and the coarser sands of the lower eastern horn; both the lower stoss and lower lee display the least amount of fine skew;
- kurtosis mirrors the signature of skewness. The crest-brink is of higher leptokurtosis while both the lower lee and lower stoss approach normality; the lower stoss is less leptokurtic (often slightly platykurtic) than the lower lee.

The spring distributions are as follows:

- grain-size is highly variable on all morphologic elements, without any significant sedimentary differentiation over the dunes;
- sorting is best on the upper micro-morphologic regions and the lower eastern horns of the dunes, with the stoss slopes being relatively more poorly sorted;
- highest fine skewness occurs upon the crest-brink relative to both the lee and stoss slope;
- leptokurtosis is greatest on the crest-brink relative to the leeward and windward slopes; lowest kurtosis is found upon the base of the dunes;

6.5 Quantitative Temporal Variation

Unlike comparisons between different micro-morphologic locations, seasonal contrasts of like morphologic elements are not as readily obvious when dealing with more than one dune. Although the sediment is restricted in granulometric range, factors that influence granulometric-seasonal variability involve dune size, hence the amount of upslope wind speed amplification, field position and the influence of intradune obstacles. Variation due to sampling, both in the field and laboratory may also influence outcomes. In the case of mobile forms, the integration or destruction of a dune part way through a study, could considerably alter the sedimentological signature.

No two dunes are exactly alike. Although profile trends may be evident between micro-morphologic position, the numerical values of the trend lines may be quite dissimilar for different dunes. Therefore, sufficient and representative numbers of dunes need to be sampled to give an accurate approximation of the sedimentological signature of the population. Such data is receptive to statistical analysis. Qualitative interpretation is less efficient when dealing with multiple sample points within a given micro-morphologic partition, but is suitable for deducing probable single sample change for a specified morphologic location such as the crest. When coupled with sample numbers allowing statistical processing, both derive a strong foundation from which postulates about aeolian processes and temporal differentiation of surface sedimentology can be deduced.

Although some clear distinctions of inter-seasonal variation for each granulometric moment are evident from the longitudinal profiles and median values of each morphologic location, many more inter-seasonal contrasts are ambiguous. Graphical perusal is not adequate for explicitly identifying temporal change, while the narrow range for numerous median values poses doubt upon the validity of the numeric differences. It is paramount that statistical verification is used to simplify and to confirm or deny these ambiguities. Table 6.3 illustrates the significance of the variation of like morphologic components for different seasons. It is apparent that differences between like elements in contrary seasons are fewer than for differences between different elements in the same season. This demonstrates that greater variability is caused by micro-geomorphologic location than by seasonal changes in the wind regime.

Comparative statistics for variation of like morphologic components for different seasons demonstrates that no seasonal differences are found for mean grain-sizes or skewness. Significant, however, are the differences for crest-brink sorting between:

- winter and summer
- summer and spring

The series of relative seasonal sorting between the end-members of better and poorer being, is revealed to be (winter/spring, autumn, summer). Sorting for winter and spring is indistinguishable and are both slightly better sorted than the autumn crest-brink region, and significantly better sorted than the summer crest-brink. Consequently, no significant differences are found between autumn, winter and spring, nor autumn and summer. As for all such series, these differences may or may not be statistically significant depending on the range between end-members.

Likewise the mean grain-size and skewness of the low lee position are not different in any one season. However, sorting is shown to differ at the 0.01 level of significance between:

- summer and spring

The relative series of sorting is (summer, autumn/winter, spring), whereby the degree of sorting for summer and autumn show 'no difference' as do winter and spring. Both autumn and winter therefore lie between the clearly differentiated surface sedimentology of the summer and spring end-members.

The comparisons of seasonal variation on the lower stoss slope demonstrate that significant differences of mean grain-size occur between:

- winter and summer
- winter and autumn
- spring and autumn

Geomorphic Element	Moment	H statistic	Significance Level	Scheffé Projection (k-groups)
crest-brink	1	2.72	not significant	no differences
crest-brink	2	18.88	0.01	(1&2) (2&4)
crest-brink	3	5.90	not significant	no differences
low lee	1	4.05	not significant	no differences
low lee	2	11.74	0.01	(2&4)
low lee	3	5.81	not significant	no differences
low stoss	1	25.27	0.01	(1&2) (1&3) (3&4)
low stoss	2	10.48	0.05	(1&2)
low stoss	3	0.95	not significant	no differences
low horn west	1	2.36	not significant	no differences
low horn west	2	5.65	not significant	no differences
low horn west	3	4.86	not significant	no differences
low horn east	1	5.32	not significant	no differences
low horn east	2	9.79	0.05	(1&3)
low horn east	3	3.52	not significant	no differences

Table 6.3 Seasonal comparisons of significant differences between different morphologic elements where: *df=3* and *k=(winter = 1, summer = 2, autumn = 3, spring = 4)*. For a significant difference to occur and to be tested using the Scheffé projection method, H ≥ critical value. Critical values being $\chi^2_{0.05}=7.82$ and $\chi^2_{0.01}=11.34$

The comparisons of the second moment of the stoss slope during different seasons confirms significant differences between:

- winter and summer at the 0.5 level of significance.

No difference for skewness was demonstrated for the lower stoss slopes. The relative series of finer to coarser mean grain-size on the lower stoss slope is given by, (winter, spring, summer, autumn), and for sorting, (winter, autumn/spring, summer), with winter being better sorted than summer.

Except for sorting on the lower eastern horn in winter and autumn, the horns of the dunes in all other seasons, for all moments, were characteristically shown as having 'no difference'. The relatively low degree of difference ($H = 9.79$) with this single exception is more allied to the non-significant differences generally found for the horns.

6.6 Summary of Temporal Sedimentological Change

6.6.1 Crest-Brink Zone

It is concluded from statistical and interpretative comparisons of seasonal crestal elements that the variability of mean grain-size for the crests of the Strzelecki dunes is not significantly different from one season to another. This does not imply that the upper morphologic position is the coarsest component over an annual cycle, but rather that, the amount of variability between samples on the upper lee, crest and upper stoss as a sample group (k-group) does not differ significantly. Conversely, the single crestal sample at the highest point of elevation does demonstrate temporal differences and may indicate changes incurred by the maximum upslope amplification of wind speed afforded by the crest. Such limited areal change of sedimentology is plausible in view of the relatively low wind energies available for sediment transport across the Strzelecki Desert overall. The relatively static mean grain-size for the dunes, as a whole, suggests that surface sedimentology is a product of the seasonal accumulation and removal of fines from over the dunes. Further discussion of these processes follows in Chapter 7.

Sorting upon the crest-brink has shown a definite temporal change, whereby crests are better sorted in winter and spring while being less well sorted in summer. Because no notable distinction occurs for the sorting of the crestal element of other seasonal combinations (Table 6.2), the differences expressed for summer versus winter and summer versus spring show these seasons to be sedimentological end-members or opposites that are capable of being identified by the Scheffé projection. To account for the non-distinction of other combinations, crestal sorting for autumn must lie intermediate between the narrow range of the two end-members of summer and (winter and spring), as well as being indistinct from each of these end-members. This is characteristic of the narrow range of values attributed to these aeolian sediments. The greater the range of sorting, the more likely it is that other combinations will show *k*-group variability; the smaller the range, the less *k*-group variability. For this reason, no differences between seasons were found for the crestal grain-size and skewness of the dunes, and it explains the limited differences in the sorting of the crestal component.

6.6.2 Lower Lee Flank

The lower lee flank reflects the absence of a statistically significant grain-size variation between seasons, but demonstrates the lower lee to be better sorted in summer than in spring. This sedimentological character may be attributed to the prevalent oblique high intensity vernal air flow across the lee slope compared to the predominant longitudinal flow recognised in summer. The absence of definitive differences between other seasonal combinations for sorting suggests that a confined range of sorting values exists, and that summer and spring represent end-member components. Any variability between winter and autumn would not be discernible, as they are near identical.

6.6.3 Lower Stoss Flank

Inter-seasonal comparisons of the low stoss demonstrate significant differences for the first and second moments. Seasonal variability of grain-size exists on the stoss slope. Grain-size is characterised as being significantly finer in winter than for both summer and autumn, with autumn being relatively finer than the stoss of spring. However, the absence of difference between many inter-group comparisons reveals similarity of stoss slope fineness for winter and spring, while summer and autumn, and summer and spring, are similar without there being any real difference in grain-sizes.

Sorting upon the lower stoss is significantly different only between the end-members of the better sorted winter and least well sorted summer sands. No clear distinction between the moment values of autumn and spring is found. Skewness did not show inter-seasonal contrast.

6.6.4 Dune Horns

Except for the difference expressed for sorting between the eastern horns in winter and autumn, both the eastern and western horns unequivocally demonstrate an absence of sedimentological seasonal differences. Samples of the winter dunes show that they are better sorted on the lower eastern horn than on the lower eastern horn of autumn. They could be inferred to be the end-members of sorting, with summer and spring values being close statistically and undifferentiated. Even with this one exception of sorting between the eastern horns in winter and autumn, the difference is the smallest found, as measured by the size of the *H* statistic (H = 9.79) away from the established chi-square value $\chi^2_{0.05} = 7.82$. This difference may be indicative of oblique winds crossing the dunes, as identified in the southern Simpson Desert by Brookfield (1970) and Fryberger (1979). Oblique air flow has also been demonstrated from the alignment of sedimentary features in the Gurra Gurra dunes, from which eastern slipfaces, ripple migration, scour marks and horn elongation have been identified (see Section 4.1.1, Chapter 4). Overall, west-to-east is the prevalent oblique direction of transport, and the contrary differences of winter and autumn may be due to different degrees of intergranular sorting under different wind intensities from the west.

6.7 Bivariate Comparisons of Lee and Stoss Moments

Bivariate associations using correlation analysis describe the direction and character of the

Season/Morphology	Moment Bivariate	Spearman Coefficient	Significance Level
winter/lee	1 & 2	0.321	not significant
winter/lee	1 & 3	-0.560	0.01
winter/lee	1 & 4	-0.701	0.01
winter/lee	2 & 3	-0.610	0.01
winter/lee	2 & 4	-0.663	0.01
winter/lee	3 & 4	0.802	0.01
winter/stoss	1 & 2	-0.123	not significant
winter/stoss	1 & 3	-0.534	0.01
winter/stoss	1 & 4	-0.313	not significant
winter/stoss	2 & 3	-0.306	not significant
winter/stoss	2 & 4	0.343	not significant
winter/stoss	3 & 4	0.805	0.01

Table 6.4 Bivariate relationships between moment measures on the lee *(n = 24)* and stoss *(n = 31)* slopes in winter, where: 1 = mean grain-size, 2 = sorting, 3 = skewness and 4 = kurtosis. For the Spearman correlation coefficient (r) to be significant, r > critical value. Critical values are *0.05 = 0.407* and *0.01 = 0.521 (n = 24)* and *0.05 = 0.356* and *0.01 = 0.459 (n = 31)* (two- tailed).

Season/Morphology	Moment Bivariate	Spearman Coefficient	Significance Level
summer/lee	1 & 2	-0.720	0.05
summer/lee	1 & 3	-0.524	not significant
summer/lee	1 & 4	-0.424	not significant
summer/lee	2 & 3	0.607	not significant
summer/lee	2 & 4	0.443	not significant
summer/lee	3 & 4	0.717	0.05
summer/stoss	1 & 2	0.011	not significant
summer/stoss	1 & 3	-0.165	not significant
summer/stoss	1 & 4	0.031	not significant
summer/stoss	2 & 3	-0.640	0.01
summer/stoss	2 & 4	-0.250	not significant
summer/stoss	3 & 4	0.097	not significant

Table 6.5 Bivariate relationships between moment measures on the lee *(n = 11)* and stoss *(n = 19)* slopes in summer, where: 1 = mean grain-size, 2 = sorting, 3 = skewness and 4 = kurtosis. For the Spearman correlation coefficient (r) to be significant, r > critical value. Critical values are *0.05 = 0.618* and *0.01 = 0.755 (n = 11)* and *0.05 = 0.460* and *0.01 = 0.584 (n = 19)* (two-tailed).

relationship between two variables. Table 6.4 illustrates the relationships between two moment variables for the lee and stoss flanks in winter. A zero correlation is achieved between mean grain-size and sorting for samples from the lee flank. However, as mean grain-size decreases so do fine skewness and lepto-kurtosis. Better sorting leads to a higher skewness and a higher degree of leptokurtosis. There is a prominent association of skewness and kurtosis; as one increases or decreases the other follows suit.

As with the lee slope data, no correlation exists between the degree of sorting and mean grain-size upon the stoss. Stoss slope samples exhibit zero correlations between the bivariate comparisons of mean grain-size and kurtosis, sorting and skewness, and sorting and kurtosis. Most bivariate comparisons of sedimentological moments on the stoss show little correlation with each other, for winter derived samples. Exceptions to this are associations of decreased mean grain-sizes (fining) with decreases in positive skewness, and lower skewness with lower lepto-kurtosity.

Season/Morphology	Moment Bivariate	Spearman Coefficient	Significance Level
autumn/lee	1 & 2	-0.945	0.01
autumn/lee	1 & 3	0.446	not significant
autumn/lee	1 & 4	0.515	not significant
autumn/lee	2 & 3	-0.287	not significant
autumn/lee	2 & 4	-0.373	not significant
autumn/lee	3 & 4	0.994	0.01
autumn/stoss	1 & 2	-0.354	not significant
autumn/stoss	1 & 3	0.501	not significant
autumn/stoss	1 & 4	-0.593	0.05
autumn/stoss	2 & 3	-0.737	0.01
autumn/stoss	2 & 4	-0.807	0.01
autumn/stoss	3 & 4	0.822	0.01

Table 6.6 Bivariate relationships between moment measures on the lee *(n = 8)* and stoss *(n = 15)* slopes in autumn, where: 1 = mean grain-size, 2 = sorting, 3 = skewness and 4 = kurtosis. For the Spearman correlation coefficient (r) to be significant, r > critical value. Critical values are *0.05 = 0.738* and *0.01 = 0.833 for n = 8,* and *0.05 = 0.521* and *0.01 = 0.654 where n = 15* (two-tailed).

The relationships between two moment variables for summer are illustrated in Table 6.5. The bivariate associations of the summer lee slope samples are generally insignificant. Only when mean grain-size fines is sorting seen to improve, while a low skewness correlates with a lessened kurtosis. Similarly, the summer stoss samples are generally uncorrelated. The only significant correlation is between sorting and skewness, with better sorting leading to a higher degree of skew.

Table 6.6 demonstrates the bivariate comparisons of the sedimentological moments for autumn. The lee slope results mirror those of summer, and only signify compatible differences between mean grain-size and sorting, and skewness and kurtosis. As mean grain-size decreases sorting improves, while a decrease in skewness is associated with a decrease in kurtosis. The stoss slope, however, shows a greater number of bi-parameter correlations. Finer mean grain-size is correlated with increased lepto-kurtosity, while better sorting is associated with higher skew and higher kurtosis. Not surprisingly, the lower the skew, the lower the kurtosis. Zero correlations are found between mean grain-size and sorting, and mean grain-size and skewness.

Season/Morphology	Moment Bivariate	Spearman Coefficient	Significance Level
spring/lee	1 & 2	0.028	not significant
spring/lee	1 & 3	-0.509	0.05
spring/lee	1 & 4	-0.377	not significant
spring/lee	2 & 3	-0.227	not significant
spring/lee	2 & 4	-0.699	0.01
spring/lee	3 & 4	0.743	0.01
spring/stoss	1 & 2	-0.367	0.05
spring/stoss	1 & 3	-0.28	not significant
spring/stoss	1 & 4	0.169	not significant
spring/stoss	2 & 3	-0.540	0.01
spring/stoss	2 & 4	-0.730	0.01
spring/stoss	3 & 4	0.880	0.01

Table 6.7 Bivariate relationships between moment measures on the lee *(n = 11)* and stoss *(n = 19)* slopes in spring, where: 1 = mean grain-size, 2 = sorting, 3 = skewness and 4 = kurtosis. For the Spearman correlation coefficient (r) to be significant, r > critical value. Critical values are *0.05 = 0.618* and *0.01 = 0.755* *(n = 11)* and *0.05 = 0.460* and *0.01 = 0.584* *(n = 19)* (two-tailed).

Bivariate correlation of sedimentological moments for spring is shown in Table 6.7. Lee flank samples from the crescentic dunes demonstrate a significant correlation between mean grain-size and skewness where diminished grain-size develops a lessened skewness. Zero correlations occur between mean grain-size and skewness, mean grain-size and kurtosis, and sorting and skewness. Both skewness and kurtosity proportion themselves as in all previous seasons. The stoss flank demonstrates that with concurrent fining of grain-size, sorting improves, and that with better sorting, greater positive skewness and kurtosis develops. Again, skewness and kurtosis demonstrate proportional responses to incremental change for both lee and stoss elements.

6.7 Overview of Crescentic Dune Surficial Sedimentology

Transitional temporal variation in wind pattern and the associated change of dune shape, is fundamental in producing the sedimentological distributions. As a consequence of seasonal morphologic variation, continual form and gradient changes lead to the ongoing reworking and redistribution of dune sands, especially within the crest-brink zone and with the finer sand fractions. Generally, sorting occurs with sand movement upslope of the stoss flank. However, multi-directionality in the wind regime is fundamental to forming the seasonal sedimentary signatures. Importantly, sand transport is not continuously upslope of the stoss for all periods of the year. In winter there was a reversed air flow pattern down the stoss flank, while in both autumn and spring air flow was oblique and easterly across the dunes. Only summer winds were dominantly parallel to the axis of dune advance, the season when sorting was at its worst for the crestal-upper morphologic positions. In view of the intensity of the seasonal wind regimes, it is not unreasonable to expect that during the times of strongest winds, crestal sorting is poor, and is due to the entire dune being active. Conversely, only the upper morphologic positions reach sustainable threshold shear velocities in the quiescent and low wind energies of winter, autumn and, in some respects, spring. Discernible zones of gradational grain-size differences are identified by the irrefutable variation in the degree of sorting on the crest-brink in winter and summer. Several inter-seasonal differences of mean grain-size and sorting on the lower stoss slope are also prime examples of this interactive spatial and temporal variation. In general, the surficial sedimentology of the Gurra Gurra crescentic dunes can be summarised as an association of finer mean grain-size with low skewness, lower lepto-kurtosis and better sorting throughout the yearly cycle. This is a finding which also agrees with other studies across the globe such as those by Lancaster (1982a, 1987); McKee (1983) and Vincent (1984).

Clearly, it has been demonstrated that mean grain-size, sorting and skewness of desert sands for the Gurra Gurra crescentic dunes, as well as the shape of the dunes, are dependent upon the temporal frame of reference. Both the sedimentology and morphology have been revealed to be cyclic in the present environmental and climatic setting, thus allowing for the prediction of moment distributions for particular morphologic and temporal settings. Such predictions are founded upon an understanding of the surficial processes over dunes and are examined in Chapter 7.

7 Crescentic Dune Morpho-Genesis and Dynamics

7.1 Dune Initiation

Bagnold (1941, p.72, 169 - 174) identified that as wind encountered change in surface roughness between a pebble pavement and a sand patch, the saltation curtain undergoes deposition resulting in the initiation of a dune. 'A given wind can drive sand over a hard immobile surface at a considerably greater rate than is allowed by the loose sand surface' explaining why sand accumulates on areas already sand covered (Bagnold, p.72). Similarly, Kocurek *et al.* (1992) determined that during dune nucleation on Padre Island, Texas, upwind speeds were reduced by changes in aerodynamic roughness or micro-topography such as erosional depressions, relict dune topography, nebkha or randomly vegetated surfaces. The crestal realm of the Strzelecki linear dunes display most of these features throughout different seasons.

The smooth clay intradune area in winter, is a relatively flawless surface that is not responsive to dune formation nor vertical down-wearing. However, once desiccation fractures develop in the hotter months, increased surface roughness of the intradune area assists in developing areas of undulation where sand accumulation may occur within deflation hollows. As a consequence of surface desiccation, an increased surface roughness transpires, which in turn leads to the accumulation of sand piles that catalyse dune initiation. The greater intensities of the winds and temperatures of summer, and to a lesser extent, spring, coincide and contribute to dewatering of the moistened sands of winter. From this, the process of desiccation can proceed rapidly and releases a seasonally periodic and plentiful supply of sediment for the evolving downwind crescentic dunes. The development of downwind transverse dune ridges, as well as the limited extent of desiccated intradune in this region of the field, testify to the upwind southern source of sand supply for the growth of the northern transverse dunes, as well as the mechanism for sediment production.

The crescentic dunes at Gurra Gurra waterhole are sited in a localised and discriminatory topographic setting that generates wind velocities that are higher than surrounding localities and are specific to the crescentic dune site. Crescentic dunes occur on a linear dune of dimensions approximately twice those for other linear dunes in the immediate area, and which also acts as an abundant supply of sand for dune building. The channel of Strzelecki Creek rises several metres from its lowest point to the surrounding plain upon which several linear dunes extend. However, no distinct differences in elevation occur between the crescentic dune site and neighbouring linear dunes, which also abut Strzelecki Creek channel. Once sourced from desiccated intradune areas, the form of the crescentic dunes is related to evolution upon relatively unconsolidated, thick substrate materials of the linear dune, and seasonally multi-directional winds. These advance and retrogress the entire dune body, and elongate the eastern horn or linguoid.

The crescentic and asymmetrical profile of the Gurra Gurra dunes is due to dominant uni-directional summer winds. As the nascent dune grows vertically from a sand patch into a dune-form, the upper elements are always subjected to higher velocity winds than are the lower elements. This process induces greater migration of the upper sections of the dunes relative to the lower, which, in turn, produces an asymmetrical shape. Shear is greatest upon the quasi-equilibrated crest, after which lee slope deposition occurs via saltation over the crest, or grain-fall and avalanche under less vibrant winds, or when slope instabilities are pronounced. Erosion of the windward slope and re-deposition of these sands upon the lee, maintains profile asymmetry and forward movement of the dune.

7.1.1 Dune Shape

The processes of erosion and deposition are fundamental to the initiation of aeolian landforms, but are also responsible for landform equilibration, by feedback mechanisms responding to changes in the environment. The equilibrium morphology is controlled by the pattern of u_*, for it is u_* that determines the rates of sediment transport. Dynamic equilibrium is the penultimate form of these feedback mechanisms. For example, the yardang obtains a geometric equilibrium when an inverted keel shape and a width-to-length-to-height ratio of 2:8:1 is achieved. Similarly, a barchan dune is in equilibrium when its dimensional ratios are *c.* 1:1.5 for width-to-length and 1:10 for height-to-width. Such dimensions portray a slipface at the angle of repose, a stoss slope no greater than 15° in gradient and a coincident crest-brink. Such equilibria for these landforms is usually obtained only in very specific environmental niches of intense aridity, where winds are prevalently uni-directional with very insignificant directional variability and obstruction to flow.

According to Burkinshaw *et al.* (1993) the morphological equilibrium for transverse dunes is a profile asymmetry that possesses a stoss slope of 10-15° and a lee gradient of 33°. Working on the reversing coastal dunes of Alexandria, South Africa, Burkinshaw *et al.* (1993) identified enhanced upslope streamline convergence at the start of a seasonal reversal due to the high angular differential for the slope-interdune boundary. This is when the dune is in a most un-equilibrated form. A dune is in a constant feedback relationship between seasonal wind direction and form as it tends towards an equilibrium shape (Burkinshaw *et al.* 1993). In this respect, the Gurra Gurra dunes achieve equilibrium only in the summer months. This finding is in agreement with the observations of Burkinshaw *et al.* (1993) for the Alexandria coastal dunes. However, the dunes of Gurra Gurra waterhole reside in an environment of multiple aeolian influences, and therefore the geometric character of the landform is one of predominant quasi-equilibrium. The following discussion outlines the quasi-equilibrium controls influencing the crescentic dunes of Gurra Gurra.

7.1.2 ENSO and Dune Activity?

The apparent cyclical stages of juvenile-to-mature evolution expressed by the Gurra Gurra crescentic dunes may coincide with the climatic variability of the El Niño Southern Oscillation (ENSO). The onset of ENSO-induced drought exacerbates aridity which, in turn, may locally increase the sand drift potential of the winds across dune crests in the central deserts. It is most significant that the time of optimum dunefield development has occurred within the time of the prolonged El Niño event that commenced in 1990-91 (Pearce, 1993). Conversely, the apparent genesis of crescentic dunes equates with the most intense El Niño event for a century, in 1982-83 (Pearce, 1994). However, even under such proposed increases in aridity, windiness does not alter significantly, for, not all dunes are reactivated. Similarly, Wasson (1976) recognised that the dunes of the semi-arid Belarabon region of NSW did not reactivate in times of severe drought and grazing in the late 1960's. Hence, increased aridity and loss of vegetative cover alone, do not seem to promote dune reactivation. The onset of ENSO not only may reduce botanical populations, but more importantly, diminish floral obstruction and retardation of air flow over the dune body. This would enhance a large dune's activity by offering unobstructed upslope wind speed amplification and hence the dune's ability to evolve to a more complex morphology. Ash and Wasson (1983) conclude that it is not dune stabilisation by plants that generally halts dune construction and development, but rather the retarding effect that plant communities present to a wind regime of existing low sand drift potential.

Evidence for this speculation is supported by the effects of autumnal plant proliferation at Gurra Gurra waterhole. Although median annual rainfall is < 150 mm for the NE Deserts (Allan, 1990), soil moisture retention is supported at Gurra Gurra, by clays interspersed as discrete illuviated laminae in the stratified sands of the linear dune. The expansion and contraction of the clay horizons form characteristic desiccation polygons on the intradune floor in summer, which quickly re-establish floral populations. These do not appear necessarily to anchor the dunes, except for mats of succulent plants upon the dune toe. The autumn field season clearly demonstrated the rapid establishment of shrubs upon the intra-dunefield, and its catastrophic effects upon the morphology of barchan and barchanoid dunes at the southern end of the field. The ragged appearance of the dunes was a direct consequence of airflow obstruction augmented by shrub vegetation such as *Cynanchum floribundum* and the bramble *Enchylaena tomentosa* (the native tomato). Although not quantitatively assessed, observations of the ragged and disturbed appearance of the autumnal dunes at the southern end of the Gurra Gurra field clearly reflected the interaction between air flow and shrub size and shape. The smaller dunes (height *c.* 1 to 1.5 m) were almost destroyed by the interference of intradune vegetation that measured only 0.5 to 1.0 m in height, but which covered more than 30 per-cent of the intradune spacing between the lee and stoss slope of the dunes. Foliage density was maintained from the floor of the intradune to the top of the plant, so that surface shear was considerably obstructed and retarded. Vegetation height and foliage density were seen to be variables that affected the rate of degradation. Even when the plant was dead, but retained a significant intrusion into the boundary layer, obstruction to air flow was sustained. Similar findings have been noted by Ash and Wasson (1983) who reported that wind speeds directly behind an equidimensional dense shrub on a reasonably flat sandy surface were reduced to 20% of the upwind value. The length of the wake of wind speed deceleration behind the shrub was controlled by shrub height and diameter and expressed as the ratio *(d/h)*. The greater the *(d/h)*, the greater the length of the deceleration area. Hence, although only speculative, it is feasible that periods of intense El Niño over the last decade may correlate with the morphologic evolution and hence maturity of the current crescentic dunefield of Gurra Gurra. Furthermore, the effects of the intensity and duration of ENSO cycles on regional rainfall and hence botanical abundance, may be responsible for cyclical phases of activity and inactivity over large linear dunes in the northeastern deserts of Australia. Significantly, no change occurs within the swales, whether vegetated or not, due to the lack of windiness for sand entrainment at ground-level. In similar fashion, Allan (1988, 1990) identified an association of 'anti-ENSO' and major fillings of Lake Eyre in the Simpson Desert and the regional climatic change that such phenomena appear to support.

7.1.3 Upslope Wind Speed Amplification

The development of crescentic dunes at the Gurra Gurra locality involve processes that are affected by the linear dune upon which the crescentic dunes occur. Evolution of the crescentic dunes into a minor but distinct dunefield appears to be founded upon the size of the underlying linear dune (Wasson, 1995 *pers.comm.*). The dimensions of the linear dune, relative to others in the area, is about two-fold, and is the only one affected by active dune building. The fractional speed-up ratio (see below and Table 7.1) is greater on larger dunes, where shear velocities over the flank(s) become significant enough to deflate and reconstruct the linear dune into a field of crescentic forms. For this to occur, it is vital that there is an increase in shear velocity associated with upslope streamline convergence of air flow between the interdune corridor and the eastern plinth of the linear dune. Plinth gradient is steeper on the east than on the west and increases fractional speed up by the greater streamline convergence and gradient contrast on the eastern side of the dune. The increase in u_*, and hence the increased sand drift potential on the eastern flank of this large linear dune at Gurra Gurra, exceeds the ambient threshold friction velocities necessary for the erosion of sand from the desiccated crest. As found by Lancaster (1989, p. 105) for linear dunes in the Namib, '"..varying wind velocities and directions over a period of time give rise to potential sand transport rates in crestal areas which are 8 - 20 times greater than on adjacent plinths."

The crestal area of the larger dune is also important in that it has the space for crescentic form evolution, when compared to the sharp, peaked and smaller linear varieties (Wasson, 1995 *pers.comm.*), where transient crestal barchanoid slipfaces often occur. Therefore, it is the combination of smaller dimensions of the surrounding linear dunes, low regional windiness and, to some limited extent, vegetative stabilisation that inhibits their reactivation into sand conveyors. The larger size of the Gurra Gurra linear dune, and its effect on increasing threshold friction speeds overcomes these restrictions. Similarly, and in support of this hypothesis, the red-brown Merty Merty linear dunes some 50 km NNE of the crescentic locality, are dunes of regional extent and significant dimensions that maintain an active sharp sinuous crested linear morphology. A transient complex linear form was observed in both summer and autumn when barchanoid dunes dominated the upper western plinth and crest-line. These dunes are, however, only approximately one-half the width of the widest structures, when compared with the width of the Gurra Gurra linear dune. The Merty Merty dunes are, however, capable of greater sand movement via upslope wind speed amplification due to their greater height. Therefore, it is apparent that dune size is fundamental in the maintenance of aeolian activity when regional potential sand drift is low, as occurs in the central Australian deserts.

It was evident from field observations of sand mobility by ripple motions, that the crestal region of a crescentic dune also undergoes the greatest changes due to increased upslope wind velocities. At the Gurra Gurra site during low wind velocities, which is for most of the day in winter and autumn, only ripple migration and minor saltation were observed on crestal areas, while the lower stoss and lower lee flanks were conspicuously inactive. This observation has been quantified by Ash and Wasson (1983) for lunettes and transverse source bordering dunes in the Simpson and Strzelecki Deserts as well as in south-western New South Wales. Buckley (1979), in the Simpson Desert, also estimated sand movement of the upper dune to be two orders of magnitude greater than on the lower slopes. Using the formula of Jackson and Hunt (1975) the Gurra Gurra dunes show a magnitude of upslope wind speed amplification lying between 0.52 for low dunes and 0.72 for tall dunes during normal winds. Under the influence of reversed winds over the slipface, amplification factors are 0.46 on small dunes and as much as 2.35 on larger structures (Table 7.1). It is in this manner, that the reversed form of the Gurra Gurra crescentic dunes is developed and maintained by regionally gentle northerly winter winds. Such wind speed modification increases the number of days during which potential sand drift is possible. Even in areas of low sand drift potential, the interaction of air flow with a dune will increase sand mobility at that specific locality. Once the dune obstacle has been initiated, the cycle of crescentic dune building can evolve, even under regional conditions that are generally not suitable for sand mobility and dune genesis.

Similarly, Lancaster (1985b, 1989) quantified the speed-up ratio for crescentic dunes in the Namib (u_2/u_1, where u_2 is the velocity at the dune crest and u_1 is the velocity at the upwind base of the dune), as ranging between 1.3 - 1.46 for isolated dunes and 1.79 - 2.25 for paired dunes. In short, this means that the larger and higher the dune, the greater is the increase in upslope velocity. The greater the velocity, the greater the sand drift potential.

This correlates with the well developed 4 - 6 metre high barchanoid and transverse ridges at the northern end of the Strzelecki site, where over the last decade, as identified on aerial

l_{stoss} (m)	l_{lee} (m)	h (m)	Amplification Factor (2h/l)	Dune Nº:Season
-	6.0	3.0	1.00	1 winter
-	9.1	4.8	1.05	2 winter
-	6.5	1.5	0.46	3 winter
-	8.4	3.3	0.79	4 winter
-	5.8	3.8	1.31	6 winter
10.8	-	2.9	0.52	1 summer
17.4	-	5.4	0.62	2 summer
12.2	-	3.6	0.59	3 large summer
05.6	-	1.9	0.68	3 small summer
10.6	-	3.8	0.72	6 summer
-	2.4	2.6	2.17	3 autumn
-	3.0	3.8	2.53	7 autumn
-	3.6	4.3	0.90	8 autumn
-	4.2	3.54	1.67	2 spring
-	2.3	2.7	2.35	7 spring
-	5.8	4.0	1.38	8 spring
-	5.3	4.6	1.74	9 spring
-	7.6	4.9	1.29	10 spring

Table 7.1 Upslope wind speed amplification for the Gurra Gurra crescentic dunes, where l is the length of the dune measured parallel to the wind at *(h/2)* and h is the height of the dune. Note that lee slope amplification occurs under the influence of reversed winds. Due to u_* being significant for the entire dune in summer, upslope wind speed modification over the stoss is a negligible influence with the existing high velocity winds.

photographs, significant development in dune sizes and numbers has occurred. As the dunes grew in size, the sand drift potential increased via upslope wind enhancement, which perpetuated and increased dune size. As the transverse dunes grew, the impact of vegetation on flow obstruction and dune morphology was significantly diminished, whilst seasons of very low wind intensity could offer some sand drift potential due to slope effects. In support of this, Table (7.1) clearly shows the greater magnitude of wind speed amplification for the larger paired or barchanoid dunes (1, 2, 6, 7), and hence the incremental ability to transport sand over the dune. The reversed morphology of separated crest-brink for the autumn dunes only prevailed on the larger dunes (e.g. N° 7), while smaller structures only reversed once the wind patterns approached the winter months of consistent reversed flow. Hence, even at the time of least aeolian activity, dune size and upslope wind speed amplification were seen to be significant at sculpturing dune forms.

Even under extremely light winds, the dunes of Gurra Gurra continue to develop due to an evolving morpho-dynamic response to the existing low intensity aeolian regime. The morpho-dynamic system once started, is self-perpetuating. As stated by Ash and Wasson (1983, p.20), the fractional speed-up of air flow upslope towards the crest and the decrease of velocity on lee slopes is an "effect that is of fundamental importance in the formation and shape of dunes because it partly determines the mobility of sand". Up-slope acceleration implies that winds not capable of significant sand movement on the lower slope will cause upper flank-crestal sand movement. Although the regional wind speed may not be significant at inducing saltation, the interaction of air flow and dune will instigate upslope acceleration so that sand movement is significant and the crestal morphology is altered, especially during winds that are not from the annual prevailing direction. This is the model that best explains the occurrence and morphological variations of the evolving crescentic dunes at Gurra Gurra waterhole.

7.2 Dune Morphology and Morphometry

It has been manifested that the morphology and dimensional ratios between the morphometric parameters of the Gurra Gurra dunes are locality dependent. Also, significant deviations of variable dependency, and hence, the adjustments of one variable in response to another, for the Gurra Gurra, Peruvian and Salton dunefields, vary considerably with geographical location. The dimensional ratios for each of the Gurra Gurra, Pampa de la Joya and Salton Sand Sea dunes clearly identify significant differences in some ratios and, hence, gross morphology, but also display some geometric similarity in others. It is inferred that, under 'ideal' environmental conditions such as:

- a prevailing uni-directional wind regime;

- plentiful but not overly copious sand supply;
- low surface roughness;
- congruity of substrate;
- negligible obstruction to air flow by undulatory terrain and/or vegetation; and
- negligible sand moisture content;

the bivariate relationships of dimensional parameters for crescentic dunes equate with an equilibrated morphology. In other words, a dynamic equilibrium exists between 3D crescentic form and aeolian environment. Such a dimensional signature is here suggested to represent a benchmark from which the degree of geometrical similarity, for different sites, can be compared. The near classical form of the barchan dunes of the Pampa de la Joya and Salton Sand Sea, therefore, express dimensional equilibrium. These dunes are established in an environment where the limits of equilibrium are rarely exceeded. Environmental processes and dune form, are in balance and remain relatively static. Whereas, for dunes in a stage of quasi-equilibrium, it is here suggested that dimensional signatures not unlike those for Gurra Gurra occur. The delicate balance of form and process is disrupted beyond the defined limits of stability. The dune system responds with a variety of morphologies that are in a constant flux of change. The rapid and variable magnitude of processes affecting the Gurra Gurra dunes do not allow an equilibrium to be fully established. In mathematical terms, the variability between morphometric parameters of the Gurra Gurra dunes is greater, and expressed as positive correlation coefficients that significantly depart from $r = 1.0$, and where the F-variance ratio does not reject a H_0 of 'no explanation'.

Characteristic of the Gurra Gurra dunes are the positive trends for all bi-variate relationships, especially the statistically significant trends between width and length in summer and length and height in autumn. Such relationships for the $x - y$ and $x - z$ dimensional planes infer morphologic quasi-equilibrium. This demonstrates that an influence of one or more morphometric variables alters the overall 3D morphology, but does not necessarily alter all interrelated geometries to the same extent.

As also implied by the diminished morphological signatures of the bivariate relationships for the barchans of the Salton Sand Sea when compared to the Peruvian dunes, the Gurra Gurra dunes are crescentic forms which significantly differ from the ideal crescentic shape. Hence, the results reported here for specific sites define a genetic lineage of form versus environmental setting.

Finkel (1959, p. 616-18) reported the Pampa de la Joya region to be a broad, nearly level plain some 1,262 metres ASL. The region is "a completely arid, barren and wind-swept extension of coarse sand covered by a stable wind-erosion pavement of closely packed, fine gravel. There is absolutely no vegetation of any sort in the entire region other than small groups of trees which have been planted at each of the infrequent stations along the Mollendo-Arequipa railroad...climatologically, the area is completely without rainfall except for a rare, freak storm...the sand of the barchans was quite free of moisture [while] the regime of prevailing winds is southerly...the absence of marked seasons and the presence of strong unidirectional winds render the site ideal for the propagation of almost perfect, classical crescentic dunes." The Pampa de la Joya dunes thereby demonstrate a morpho-dynamic, equilibrated form evolving under prevailing uni-directional winds, an abundant but not excessive sand supply, a smooth, cohesive surface plain with no vegetation or notable physiographic obstacles, and negligible sand moisture. On the other hand the Gurra Gurra dunes typify the converse setting, with multi-directional seasonally-controlled winds with a single dominant wind vector that maintains a crescentic shape, a copious sand supply, an undulatory substrate, a varying surface roughness and a vegetation cover that is seasonally variable, as well as a variable sand moisture content.

In summary, two environmental-morphological end-members are portrayed that can be defined respectively as annual mono-microclimatic and annual poly-microclimatic regimes. Between these extremes lie a range of intermediate and variable morphological-environmental arenas, that encompass many and varied terrestrial and extra-terrestrial deserts and crescentic dune morphometries and morphologies. The degree of disparity from equilibrated shape will identify the presence of site specific spatial and temporal environmental inhomogeneities. For instance, such inhomogeneities or morpho-dynamic perturbations may be due to bi- or multi-modal wind directions, textural contrasts in surface roughness, variable substrate cohesiveness or moisture content, topographical setting, sand supply, grain-size and sand bed thickness *etc.* This is demonstrated by the excellent, but not perfect correlation, of the Peruvian bi-variables (Chapter 4, Section 4.2.2). Hastenrath (1987) identified a permanent horn asymmetry for the Peruvian dunes induced by intense night-time katabatic winds. Such obliquity is the most probable cause for the absence of near perfect bivariate correlation of morphometric parameters, while Long and Sharp (1964, p. 150-151) inferred that the physiographic setting of the Salton Sand Sea is responsible for horn elongation and the distortion of dune shape. The elevation in the area of the Salton dunes is some 30-230 ft BSL (*c.* 10-70 m below sea level), and the geology is weakly consolidated clay and mudstone and sandstone beds of the Plio-Pleistocene Brawley Formation. A consistent, prevailing westerly wind blows over a surface of "smooth ground sloping gently eastward and northeastward toward the Salton Sea at 40-100 feet per mile...wind blown sand is sparse between the dunes except to the lee of scattered creosote bushes and as accumulation in gullies...the ground surface between dunes has a patchy residual armour of pebbles, concretions and cemented sandstone fragments derived from the underlying beds [with a relief of scattered bedrock knobs and local gullies] that, interferes with the movement and distorts the shape of dunes crossing it." Hence the environmental setting of the Salton Sand Trough is one that deviates from the ideal setting displayed by the Arequipa plain (Pampa de la Joya) of Peru. The 3D character of the dunes of the Salton Trough reflects site idiosyncratic perturbations, that detract from the mathematical harmony of form, but still enable barchans of good form and dimensional correlation to evolve (see Chapter 4, Section 4.2.3).

The barchans of the Pampa de la Joya can be regarded as a benchmark with which the dimensional similarity of other crescentic dunes can be compared. As discussed in Chapter 3, Section 3.1.0, and by Breed and Grow (1979) and Finkel (1959), when conditions are such that both geometrical and dynamical similarity are met, complete similarity occurs between like landforms, regardless of size, lithology and geographical locality. Thus, comparisons of different dune forms at different localities generates greater understanding of the dynamic character and spectrum of environmental niches in which dunes evolve. Although the Gurra Gurra dunes are of the crescentic variety, mathematical analysis is clear evidence that distinct morphometric differences exist between these dunes and crescentic types from other terrestrial localities, as well as those from the Hellespontus and North Circumpolar regions of Mars. In comparison with the highly correlated three-dimensional relationships expressed by the dunes of Pampa de la Joya, there are no precise cause and effect relationships when dealing with the Gurra Gurra dune population. Clearly, the spectrum of crescentic morphologies and the site specific influences on the morphometric character and degree of geometric equilibrium are thus emphasised. Site specific parameters that influence the morphological signature of the Gurra Gurra crescentic dunes are further explored below.

7.3 Wind Regime and Morphological Congruity

A distinct pattern of mathematical incongruity exists for some of the bivariate relationships of the Gurra Gurra crescentic dunes. Figure 4.3a demonstrates a relationship of dependency between width and length during early-summer, which coincides with the onset of prevailing uni-directional winds. Moreover, field observations in mid-summer, also demonstrate that dune shape for individual members of the field (e.g. dune 3, see Chapter 5, Fig. 5.1) correlates with the form and morphometry typical of the Pampa de la Joya barchans.

Low correlations between morphometric measurements recorded in late spring - early summer, however, suggest the form at this time, to be inherited from the aeolian regime of spring. Oblique winds from the west having favoured the elongation of the dune horns. Hence, the morphometric relationships expressed in Figure 4.3 a-c are due to the process of seasonal horn elongation. These measurements portray a dunefield in quasi-equilibrium as revealed by the low but significant correlations and *F*-ratios of width-to-length and the relationship between the wavelength of the compound ridges and ridge length. As summer proceeds and the winds become prevalently uni-directional, the dunes approach and achieve equilibrium of form and spacing.

It is apparent that *x-y* dimensional asymmetry is caused by horn elongation towards the east, with concurrent shortening of the western limb. Ideal 3D equilibrated form is rarely achieved by the Gurra Gurra crescentic dunes, whereby dune width is distorted significantly by oblique air flow. The relationships expressed for the winds of mid-autumn and late spring (Fig. 4.3d-f), show an absence of bivariate relations when width is one of the two variables. Bivariate associations involving width do not reflect high coefficients of correlation. Conversely, the relationship between length and height is least affected by oblique transport and demonstrates a strong correlation.

Simply stated, the seasonally dependent strong multi-modal influence of summer southerly and vernal westerly wind vectors, detracts from the form produced in a prevailing uni-directional wind. The morphometry of the Gurra Gurra dunes is dominantly affected by oblique winds that elongate the eastern horns-linguoids, and shorten the western wings throughout spring and early-to-mid autumn. This process alters dune width, which places the dunes in a state of quasi-equilibrium (Fig. 7.1). The presence of eastern horn-linguoid elongation is characteristic of oblique westerly winds at Gurra Gurra waterhole. Similarly, Bagnold (1941) and Lancaster (1980; 1982a) have also observed and modelled this phenomenon in terrestrial ergs. Tsoar (1978, 1984) also noted this in respect to both terrestrial and Martian sand seas (refer to Chapter 2, 2.11). In like fashion, the development of seif linear dunes from individual barchans at Gurra Gurra waterhole has occurred. Horn elongation is oriented by weaker, but persistent westerly winds within a bimodal regime that reverses the primary vector between NNE and SSW (Fig. 7.2). The strongest secondary oblique winds are most typical of spring, which is, therefore, the most likely season during which barchan elongation occurs.

COMPARATIVE SEASONAL VARIATION IN DUNE WIDTH AND LENGTH

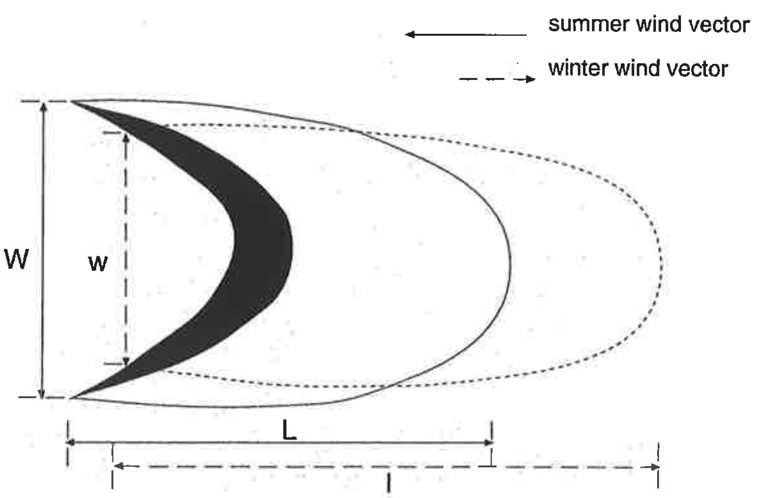

Figure 7.1 A simplified schematic diagram (not to scale) of horn retrogression by reversing winter winds which instigate a diminished width (w < W) and mass transfer of sand back over the dune causing an increase in winter dune length (dashed outline) relative to summer dune length (full outline) (l > L).

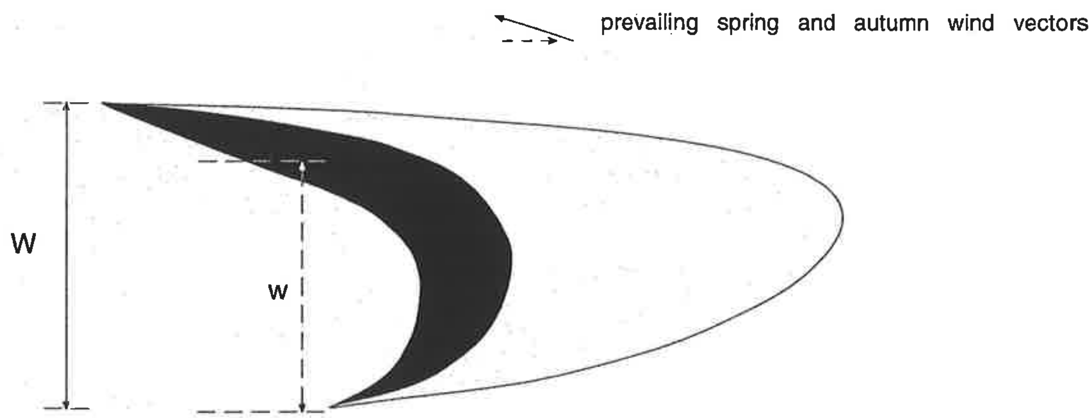

Figure 7.2 Under a bimodal regime of normal and oblique wind vectors the sum of the vector components establishes horn elongation, which in turn may exaggerate dune width (W > w) relative to a symmetrical planform that is in geometric and dynamic equilibrium (formed in prevailing unidirectional winds) (Fig. 7.1), although concurrent shortening of the opposite horn may not show a significant difference in width. When such conditions occur at Gurra Gurra, the dunes are (i) at a time of dimensional quasi-equilibrium and (ii) dune length is also distended by the action of reversing winds in both late autumn and spring. Thus a morphologic cycle of near-equilibrium and symmetry in summer transcends into autumn which inherits the features of summer, until advanced quasi-equilibrium is staged through horn retrogression, width diminution and length progression in late autumn and winter. Subsequently, both horn elongation and length progression occur in late winter into spring, but as spring wanes, the dunes travel through stages of quasi-equilibrium and return to near-equilibrium as the hotter months approach.

The model which best explains the evolution of seif-linear dunes from the barchans of Gurra Gurra, is that of Tsoar (1984) and reflects:

- barchan development under a southern primary wind in summer with a gentler secondary oblique vector stemming from the west during late spring - early summer;
- horn elongation occurring for only the eastern horn, with concurrent erosion and shortening of the western horn;
- the angle between the orientation of the longitudinal axis of the barchan and the alignment of the seif dune at 110°;
- both primary and secondary oblique winds being involved in the process of horn elongation and the maintenance of linear morphology;
- bi-directional oblique flow directed along both flanks of the seif-linear dune leading to its continued growth.

7.3.1 Dune Advance and Retrogression

Dune advance is approximately 10-14 ma^{-1} for the study dunes, and is greatest during the prevalent summer south-south-easterly wind regime. Some retrogression of dunes does occur, although the positions, sizes and forms of the dunes affect the rates of movement. Retrogression is made obvious by the recession of the horns and is characterised by a plan view on which horns of individual barchans are blunted or missing (Fig. 7.2 & 7.4). Other indicators include crestlines being somewhat more westerly, as in autumn, or elongation of the eastern horns towards the east, as prevalently observed in spring (Fig. 7.2). Howard *et al.* (1978) found that dunes in variable wind regimes were blunter than those affected by uni-directional winds, a result that is well illustrated by the Gurra Gurra dunes. The lee faces of autumnal dunes lack significant horn elements, which in turn depict a distinct leeward blunting of the dunes. Such morphologies are characteristic of oblique winds slicing the leeward slope and eroding barchan wings. Both eastern horn elongation and western horn blunting can cause minor leeward retrogression by as much as 1 m. However, greatest retrogression (3 m) occurs on the horn tips, as they also migrate laterally. Most plausibly, horn retrogression is the process most responsible for dune narrowing. The approximate gross length-to-width ratio of 1:0.4 compares with 1:0.7 and 1 :0.8 for the dunes of Pampa de la Joya and the Salton trough respectively. As both horns retrogress in late autumn and winter, and the western horn shortens in spring, the parabolic curve of the leeward section of the Gurra Gurra dune returns a variable dune width between seasons. The minimum extension of the curve and, hence, the horns is at the time of reversed winds. The greatest width occurs under normal or oblique winds, when either both horns or the eastern horn elongates. In both cases, dune width increases, and produces a dune form in quasi-equilibrium under continual adjustment to varying wind directions. Significantly, the Gurra Gurra dunes, as well as the Kelso dunes in California (Sharp, 1966), and dunes in the northern Sinai (Tsoar, 1974) showed that seasonal variations of wind direction and the seasonal construction and destruction of morphologic features, do not destroy the principal shape of the crescentic dune. A uni-directional wind, of sufficient intensity to produce crescentic dunes, influences dune shape, although during most of the year, form is influenced by other less intense winds of variable directions.

In similar fashion to the autumn form, the northern most end of the dunefield displays a lack of horns on the front row of transverse dunes abutting the fluvial channel of Strzelecki Creek. The absence of dunes within the stream channel, and the lack of dune horns on the frontal ridge, are apparently due to aeolian erosion by oblique winds; although rare high-magnitude flooding would also be periodically destructive to the transverse dune-front. Oblique winds channelled westward along the ephemeral water course and enhanced by flow past a 'teardrop' island or mid-channel leminiscate loop (Fig. 1.4, Chapter 1), may assist aeolian erosion by enhancing air-flow velocity. Although the relief of the teardrop island is subdued (< 2 m), topographic variation may be sufficient to cause air flow convergence, a venturi effect, to cut obliquely across an advancing dune-front. Lateral dispersion of dune sediment down the channel is marked by a pronounced colour change in the fluvial sediment to the west of the dunefield. The bunching of the transverse dunes as they approach the fluvial channel is indicative of an environment of abundant sand supply, where height increases more rapidly than does spacing. As the size of the transverse dunes increases, migration is retarded and successive rows of ridges begin to concertina behind each other. The smaller and rearward dunes may also act as a source of sediment to the forward ridges and assist in their growth by merging with the slower moving structures.

7.4 A Two-Dimensional Morpho-Dynamic Analysis

The equilibrated shape of the Gurra Gurra crescentic dunes, is that of a linear-linear longitudinal profile with a coincident crest-brink, a lee slope of 32° and a median stoss inclination of ~08°, formed under the regionally dominant SSE winds of summer. However, dune morphology is more often in varying states of quasi-equilibrium, with the shape and inclination of the lee to crest-brink region continuously changing under the influence of both regional and site-specific, multi-directional and variable intensity seasonal winds. In contrast to the dune shape of mid-summer-early autumn, the mid-winter-early spring form of the lee slope is typically convexo-curvilinear, terminating in a separated crest and brink, with a median slipface inclination of 17°. The lower inclination of the lee slope and the separated crest-brink morphology is typical of the actions of the more gentle and sporadic, reversed northerly winds of winter. Although the spring dunes were observed to be convexo-curvilinear, secondary westerly and oblique high velocity winds swing towards the south as the season advances and return the dunes to the morpho-dynamism of summer. Nevertheless, reversed air flow and reversed morphology is typical of the cooler seasons. To examine the dynamic processes responsible for these morphologies, the following section identifies the most likely aeolian activity affecting the lee, crest-brink and stoss of the dunes when evolving under the influence of a seasonally multi-directional wind pattern.

7.4.1 Lee Slope Morpho-Dynamics

7.4.1.1 The Reversing Lee Eddy and Intra-Dune Abrasion and Deflation

Lee slope dynamics in mid-to-late summer and early autumn conform to the concept of crestal flow separation and the production of a reversing lee eddy. Evidence of reverse air flow opposite the direction of dune migration is minimal and may be attributed to the seasonal changes in wind directions. However, characteristics that may be attributed to the existence of a reversing lee eddy are demonstrated by:

- the orientation of ripple patterns towards the SE within the dune courts in summer;
- the entrapment of dried plant materials on the lee slope when floral populations and linear-linear form were abundant in autumn;
- and possibly the deflation and erosion of the leeward intradune with consequential development of meso-yardangs.

Minor leeward erosion has been observed by Sharp (1979) and Sweet *et al.* (1988) in the Algodones Dunes, USA, and by Hunter and Richmond (1988) for dunes of the Oregon coast. Hunter and Richmond (1988), observing the daily cycle of change for the dunes on the Texas and Oregon coasts, identified erosion and deposition near the base of the lee slope. Where strong wind conditions were prevalent, they concluded that a lee eddy eroded a basal trough at the foot of the slipface. Crestal winds that were not strong enough to produce a lee eddy, were observed to be capable only of sand transport across the crest and deposition onto the lee slope. Greeley and Iversen (1985) and Iversen (1985) also reported upslope leeward separation and windward sand transport in the Kelso dunes, California. Whether such processes occur at Gurra Gurra waterhole are doubtful. Basal scour in the lee of the dunes was not observed, although the season of greatest wind energy, summer, corresponded with greatest forward migration of the dune body. Either basal scour was rapidly infilled by the migrating dune, or, dune advance inhibited leeward erosion. In either case, no evidence of a reversed lee eddy was found using these morphological criteria.

However, work by Hoyt (1966) in the Namibian Desert, SW Africa, and in the coastal dunes of Sapelo Island, USA, provides some similarity with observations made at Gurra Gurra. Hoyt (1966, p. 139) directly observed plant debris being "held in dynamic equilibrium part way up the lee face by the ascending current" of the lee eddy. According to Hoyt (1966), such eddies were seen to be effective at deflating the courts of crescentic dunes, leaving them relatively free of fine sand and forming current ripples with leeward slopes facing towards the dunes slipface. Similar observations were also recorded at Gurra Gurra waterhole. Nonetheless, these features could as easily be due to the onset of the reversing aeolian regime of winter. The presence of several nascent reversing slipfaces at the time when dead plant debris was found upon the mid-lee slope offers support for the existence of a transient reversing wind direction that would eventually intensify and persist as winter approached. Similarly, barchanoid-court deflation may also be a consequence of reversed winds sweeping sands back onto the lower lee slope. Such features are therefore more likely to be attributed to the multi-directionality of the Gurra Gurra wind patterns. In accord with this, Sharp (1966) found no evidence of a fixed or dominant lee eddy at the Kelso Dunes, California, and questioned its importance in dynamic dune shaping. Similarly, the evidence of reverse flow lee eddies on the dunes of Gurra Gurra is scant, and if present, they have a very minor role in the sculpturing of form.

Furthermore, the likelihood of a reversed eddy of air flow excavating barchanoid encapsulated yardangs is most unlikely, especially at the reduced wind velocities typical of the leeward zone of crescentic dunes. The prevalence and grouping of meso-yardangs (Chapter 4, Section 4.1.1, Fig. 4.2) between the horns of barchanoids with greatest curvature may or may-not be supportive of lee eddies and the process of lee-eddy excavation of the intradune. It is just as likely that dunes may migrate over existing yardangs and that the curvature of the barchanoid protects those yardangs in the lee court of the dune from further erosion. In contrast, yardangs outside of the dune court are more susceptible to weathering and degrade more rapidly. Even in the high wind strengths of summer at Gurra Gurra waterhole, dune courts were not perceived to be abraded by reversed flow. As observed by Hoyt (1966) in the Namib, there is clear evidence of these areas being swept clear of fine sand deposits (Fig. 7.3), resulting in a coarse lag, infilling of erosion hollows and inter-yardang corridors. However, there is negligible evidence for lee eddies having the necessary power to erode the relict linear dune from between the horns, in as much, as to excavate meso-yardangs. Higher wind energies were observed to occur along the intradune corridors than within dune lee areas at Gurra Gurra, yet here meso-yardangs are few, emphasising the rapid erosion rates of obstacles not protected between dune horns. Lancaster (1989) reported abrupt velocity diminution in the lee of crescentic dunes; the average velocity factors at the bases of the slipfaces were between 0.4 - 0.5 the velocity on the crests for paired dunes, and 0.6 to 0.8 the crestal velocity for individual barchans. According to these results and the observed decrease of lower lee activity relative to the saltating crest of the summer Gurra Gurra dunes, leeward excavation (abrasion) is most unlikely, with only minor deflation and granule ripple orientation being the most likely activity of any reversed eddy.

7.4.1.2 The Reversing Lee Eddy and Sand Apron Deposition

Evidence of reverse transport onto the base of the lee slope is provided by the absence of sand aprons in mid-summer and early autumn, but their prevalence in winter and spring. The line of intersection between the base of the slipface and the intradune corridor sharply defines a comparatively sand-free patch in summer and early autumn. However, sand-free courts are also characteristic of apron-less dunes in spring, where a curved, sand devoid (excepting for lags), leeward pattern extends some 5-10 m from the base of the slipface (Fig. 7.3). This either strongly suggests deflation via reversed flow within the dune courts, or the lack of duneward deposition from separated and forward flow of sand leaving the crest. Considering that forward motion is greatest in summer, the latter scenario is inconceivable. In the former case, the sweeping of finer particles to the base of the lee, marks a sharp intersection between intradune and dune. Due to the lee eddy being of insufficient shear strength to transport coarser grain-sizes, these particles remain as lag deposits within deflation hollows inside the dune court, and/or as short ripple segments along yardang corridors where channelling of the wind and an increase in u_* transport the larger grain-sizes. Such observations were made in the lee of the Gurra Gurra dunes and are further supported by sedimentological evidence, where the lower lee of summer and early autumn consist of fine sands and may, in part, be due to the influence of the lee eddy, as outlined above. The higher wind speeds of summer enable fine sands to be transported

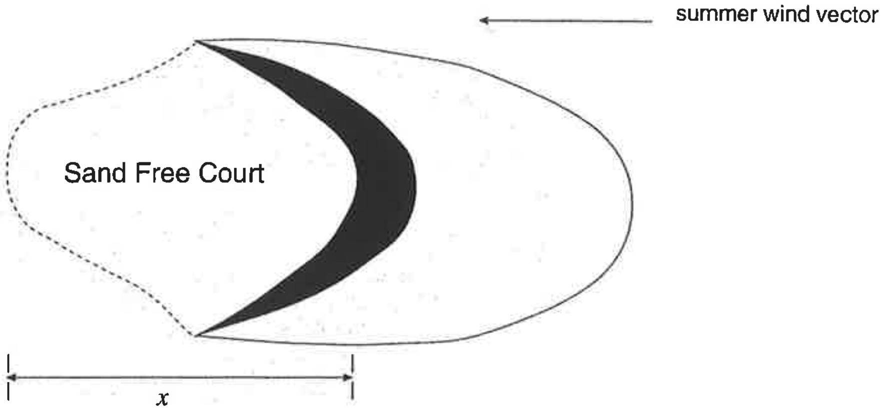

Figure 7.3 A simplified schematic plan-view diagram (not to scale) showing how the leeward courts of the Gurra Gurra dunes charateristically display 'sand-free zones' that lack any deposits excepting for coarse lags of fluviatile quartz. The morphology of the court is distinctively curved and reflects control by the planform and size of the dune, with each sand-free zone *(x)* extending some 5-10 m from the base of the slipface. This feature was readily observed in both summer and spring under the influence of strong winds, when input of fluviatile sediment and the deflation of very fine grades from the lee and intra-dune were dominant.

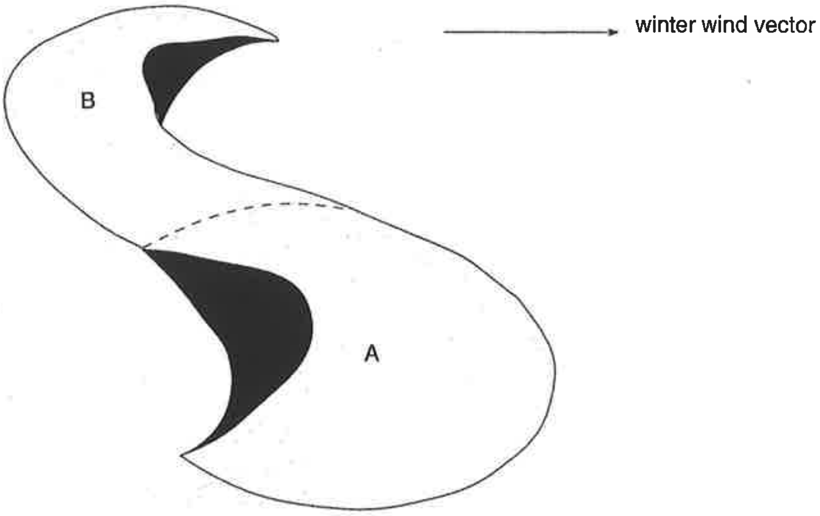

Figure 7.4 A simplified oblique view diagram (not to scale) of a complex dune-form showing (i) horn diminution and a (ii) broad domical morphology (dune A), under the influence of reversing winter winds. Slipfaces are narrow and somewhat indistinct due to dune reversal and lowering of the angle and plan-curvature of the lee slope. Once a total reversal is achieved, a 'new' slipface evolves (dune B). Such a process is, however, infrequent for the majority of Gurra Gurra dunes. This illustration demonstrates near re-equilibration of dune (B) due to the small volume of the initial sand patch (eastern horn) and hence the ease with which such masses are continuously re-sculptured in multi-directional winds.

and deposited at an acute slope angle on the lee flank by reverse eddy flow. Alternatively, the fining and concentration of fines upon the crest and lee of the summer dunes may be a consequence of fine grain overshoot under high velocity winds, with only minor reverse eddy influence. The angle of repose and substantial forward migration of the entire dune inhibits sand apron evolution. This is in contrast to winter conditions when the lower leeward inclination and static nature of the dune, with significant grainfall from a reversed and lowered lee-crest, favours sand apron formation.

The presence of winter and spring sand aprons most probably reflects downslope grainfall and erosion of the upper primary slipface. The ragged edge and low inclination (< 2°) of the sand apron is characteristic of low intensity wind gusts sweeping over the intradune and abrading, or at least winnowing, the mid-to-upper lee slope. The abrupt intersection of the intradune, and an inclination that reaches a maximum at the angle of repose for the lee slope, may be capable of causing flow separation at the base. Tsoar (1983b) has shown by experimental study of echo and climbing dunes, that air flow onto a surface inclined at < 45°, does not cause flow separation. Two-dimensional rotational flow only occurs at the windward base of a step if the slope gradient is > 60°. This indicates that scouring at the base of the slipface for the Gurra Gurra dunes and, hence, lower slope instability and random slumping onto the intradune, is unlikely as such angles far exceed the maximum angle of repose (~33°) for incohesive sand dune slopes. Contrary to these experimental 'step models', which deal with quite different wind-landform interactions, observations by Burkinshaw *et al.* (1993) have shown that lee slope undercutting and instability occur during reversed winds, due to enhanced streamline convergence between the acute angle of the interdune and dune slipface at the onset of wind reversal. Kocurek *et al.* (1992) have shown flow separation to occur on slopes exceeding 22°. Thus, some ambiguity exists between theoretical models, using solid obstacles within the boundary layer, and field observations of processes on sand dunes. The process of avalanching and the displacement of lobes of slumped sediment splaying over the intradune as an evolving sand apron in the Strzelecki, could result from displacement of the lower lee, in the manner explained above, and at angles much less than those suggested by the laboratory models. In like manner, displacement of the upper slipface could also occur, due to upslope amplification of wind velocity. Notching of the crestline, combined with fluting and slumping of slipface sands on only the upper slope of the Gurra Gurra dunes, provides evidence for this mechanism. Once the angle of the lee has been significantly lowered by reversed air flow, this process may become dominant in continued sand apron growth. Avalanching is also significant in lowering slopes and is probably an adjustment mechanism that maintains slope stability with excessive basal undercutting. During winter, lee eddies are impossible in the reversed airstream. However, it is probable that flow separation over the crest induces a smaller reversing eddy that influences the form and sedimentology of the upper stoss and crestal regions.

7.4.2 Crest-Brink Morpho-Dynamics

Crest-brink separation for the Strzelecki dunes is a process that is not confined to one model (refer to Section 2.6.3). The model of a sequential evolutionary path whereby separated crest-brink (convex) dunes evolve to dunes with coincident (straight) crest-brinks (Hastenrath, 1967; Jakel, 1980; Lancaster, 1985a, 1987) does not apply directly to the Gurra Gurra dunes, which evolve through cycles of separated and coincident crest-brink forms. An evolutionary path attests to a morphology that has become the dominant or surviving form. Neither separated nor coincident crest-brink morphologies are dominant beyond the seasonal cycle of wind changes for the Gurra Gurra dunes. The interaction of some dunes at Gurra Gurra in winter, exhibited how similar sized domical and barchan structures physically coalesced into a single dune complex. Both domical and reversed crescentic morphologies (Fig. 7.4) had evolved simultaneously. Such dual morphology is a consequence of a reversing wind, and most probably the rounding of form that exists prior to the re-development of a slipface in the presence of either a seasonally advanced reversal or restoration to the former and normal wind direction. Similarly, Bagnold (1965, p.217) made near identical observations, while Fryberger *et al.* (1984) in the Jafurah sand sea, identified an abundance of dome dunes of all sizes that were undergoing constant evolution between dome and barchan forms. The evolutionary model of crestal development is not always applicable, and cannot be considered the only process by which the more elevated elements of barchan dunes are formed and maintained. This is especially so in areas of alternating and opposing winds, as for the crescentic dunes of Gurra Gurra waterhole.

High velocity summer winds maintain a coincident crest-brink morphology. Therefore, deceleration of wind velocities incurred as seasons change results in the migration of the zones of maximum shear, erosion, and maximum deposition. Assuming that wind shear peaks at the crest for an equilibrated form (Lancaster, 1989) then peak wind acceleration would increase erosion and, hence, leeward transport over the greater length of the stoss-to-crest slope under conditions of high u_*. Under sustained conditions of this type, the Gurra Gurra dunes migrate to the NNW as the entire stoss-to-crest slope undergoes fierce saltation. Conversely, conditions of lower overall u_* would only cause erosion upon the crestal portion of the dune, due to upslope wind speed amplification. As a consequence, low shear velocities shave the upper portion of the dune with subsequent deposition onto the upper lee slope. The occurrence of crestal shaving and low velocity deposition on the mid-to-upper lee slope, reconfigures the coincident crestline to a separated convex-curvilinear feature, typical of low velocity winds as the hotter months proceed.

The extent of crest-brink separation is defined by the degree of crestal shaving, which in turn, is controlled by the wind's ability to erode and the time frame in which this process can occur without disturbance. As wind strength waxes and wanes, the degree of separation also increases or diminishes. With the Gurra Gurra site demonstrating significant seasonal wind variability, the establishment of short-term quasi-equilibrated crestal morphologies is idiosyncratic of the site. However, to further enhance this model, variable wind directions must be taken into account. The above scenario suggests the manner in which crescentic dunes in a dominant uni-directional wind, could seasonally alternate the crestal morphology between coincident and separated forms. However, the reversed and oblique nature of the Strzelecki aeolian regime, also considerably alters the form of the Gurra Gurra dunes. Only the summer form is in equilibrium, with evidence of slight separation of the crestal morphology occurring under the onset of waning inter-seasonal wind strengths, while all other seasons display varying degrees of quasi-equilibrium.

Burkinshaw *et al.* (1993) found that the area of the upslope compression cell is dependent upon wind strength, and that in seasons of light winds only the brink edge is eroded, while stronger winds erode larger regions downslope of the brink. The amplification of upslope shear velocity is reported to be approximately three times the undisturbed shear velocity (Burkinshaw *et al.* 1993). This would equate to a 30-fold increase in sand drift potential in view of the rate of sand transport being proportional to u_*^3. The winter morphology of the Gurra Gurra dunes shows a distinct separation of crest and brink with the crest having slightly greater elevation (a zone of greater deposition) combined with significantly lower shear velocities than those observed in summer. Upslope amplification of wind speed is seen to occur. When saltation is prevalent only upon the crest-brink, but absent on the lower lee and stoss, separation of the crest-brink to a convex summit by shearing the sharply demarcated and steep lee and crest of the equilibrated summer form results. According to Burkinshaw *et al.* (1993), winds that approach normal to a slope at an angle of repose (32° ± 1°), undergo extreme flow compression near the surface at the top of the slipface. Greatest shear velocity occurs at this point. As the steepness of the primary lee slope diminishes, and migration and separation of the crest towards the stoss slope proceeds, gradient contrast induces flow separation, flow deceleration and deposition. This in turn causes a feedback that further alters the gradient and the velocity profile over the dune. The high degree of crest-brink separation during winter conditions is coincident with the maximum non-equilibration. This is in agreement with the observations of Burkinshaw *et al.* (1993). If the winds were to last long enough and be of sufficient strength, a total reversal of the Gurra Gurra dunes would eventuate. The transfer of the crest over the stoss would result in the dune re-equilibrating with a coincident crest-brink and a new reversed slipface at the angle of repose. In the process of crestal migration, however, the Gurra Gurra dunes do not achieve the fully re-equilibrated form and only become quasi-equilibrated before they commence their return to near-equilibration under the summer wind regime. The presence of a rear slipface and a mobile transgressive-regressive crest is a signature of this partial reversal.

Major features of the winter winds are upslope amplifications of shear velocity, flow separation upon the crest-brink, due to the contrast between the more gentle upper slope and steeper lower slope. Concurrently, a rear slipface develops. The expression of a rear slipface combined with avalanching 'bottleneck' grain-flow, morphologically characterises the existence of reversed winds. The concordance of crest and rear slipface develops a large brink zone on the leeward side of the dune. The development of the rear slipface on the windward slope is a response to the winnowing of fine sand from the brink (upper lee) and crestal elements. This alters the surface roughness and threshold velocities of sand movement, with the contemporaneous deposition of finer fractions in a windward position upon the stoss slope. For the Kelso dunes in California, deposition on one flank is associated with erosion from the other (Sharp, 1966). The Gurra Gurra dunes show that an increase in surface roughness fosters greater deposition upon the crest, but at the same time, the upslope amplification of wind speed causes the crest to migrate. As a step inclined at 26° exists between the crest and stoss, reversed flow separation should occur over the crest, thereby assisting in the formation of a rear slipface, as prevalent for all winter dunes.

Kocurek *et al.* (1992) also identified that flow separation occurred on slopes exceeding 22°. Subsequent reattachment of the flow along the windward slope may account for the mid-windward change in grain-size experienced for many of the dunes. The winnowed fines that are an integral part of the crestal zone of the summer-autumn dunes, are redeposited on the windward slope in winter. Such a field-based hypothesis correlates with the model of Jensen and Zeman (1990), who postulated that, with seasonal change, summer wind deceleration separates the sharp summer crest-brink, with later reversed winds developing a secondary slipface. Similarly, Tsoar (1974) also associated the degree of curvature of the crest-brink, the development of a secondary slipface and the overall plan asymmetry of barchans in the northern Sinai, with two temporal stages of seasonal wind directions and magnitude. In the Strzelecki, both maintenance of a coincident crest-brink in mid-autumn for most dunes, as well as the nascent form of crest-brink separation and juvenile rear slipfaces, demonstrate the achievement of a quasi-equilibrium form at this time of year. Secondary slipfaces are formed, or at least partially developed, prior to the maximum separation of crest and brink in late autumn - early winter when reversed winds dominate.

The evolution of a separate crest-brink has produced, for some dunes, an interface of distinct gradient change upon the reversed upper lee slope in both winter and spring. This change in inclination recognises the commencement of the gentle convexity of the brink and the underlying and slightly steeper lee slope. This point may be the 'bulge' identified by Anderson (1988) where sands under the pivot point undergo avalanching as evidenced by avalanche tongues or slumps. Slight undercutting of the lee slope at this point produces a bulge. Such a process may also contribute to the accumulation of sands down the lee slope into sand aprons at the foot of the slope. The micro-morphologic development of both an upper lee pivot point and a sand apron in winter and spring, are suggested to be more than mere coincidence. Dune form is not a segregated series of components, but rather an integrated entity where change to one micro-geomorphic location instigates adjustments with other micro-geomorphic locations.

Due to the reversed winter winds being of low friction velocity, relative to the summer air pattern, the summer form is not obliterated, nor do the dunes become fully reversed. Likewise, the lower intensities of the oblique autumnal and vernal winds have little impact on the overall crescentic form of the dunes, but rather instigate characteristic seasonal morphologies. Nonetheless, these winds still undergo upslope fractional speed-up. The significance of the upslope reversed and accelerated winds of winter and spring, as well as a significant oblique vernal air stream, is displayed by the establishment of a secondary slipface between crest and stoss in both winter and spring, with a tertiary easterly facing slipface between the crest and eastern stoss flank in spring. It is also considered that such wind directions assist in the development of the overall convexity of form and, therefore, the degree of separation between the crest and brink. In support of this amendment to the hypothesis of crestal configuration, both Havholm and Kocurek (1988) and McKee (1966, 1982) deduced that crest-brink separation is a result of aeolian seasonal changes, where separation of the crest-brink decreases under the influence of cross-winds along leeward faces. Similarly, the longitudinal profiles of the vernal dunes (Fig. 4.1b), show less separation and higher slope angles (Md = 22°) than do the winter forms (median slope angle = 17°), and is inferred to be a response to the high intensity of the oblique air passage across the lee face during spring afternoons. Lancaster and Nickling (1994) have commented on the presence of relatively high wind velocities associated with oblique primary winds and low aspect dunes. Similarly, the high leeward oblique flow slicing the slipface at Gurra Gurra inhibits maximum crest-brink separation being reached in spring. Maximum separation is only achieved when the winds are reversed and normal to the crestline.

The model which best explains the collective morphological and sedimentological changes of the Gurra Gurra crescentic dunes, is that which encompasses both variable wind directions and intensities, as well as aspects of the aerodynamic hypothesis of dune development (see Chapter 2, Section 2.6). It is apparent that no single model is a satisfactorily explains the genesis, evolution and spatial and temporal variation of the Gurra Gurra dune site. A collective model will encompass the complexities of the multi-modal seasonal aeolian environmental setting of both the Strzelecki desert and Gurra Gurra waterhole locality.

The morphological transition from dome-like and barchan dune-forms, developing downwind into barchanoid and transverse forms under the principal summer regime is characteristic of a prevalent unidirectional setting. However, the interplays of winter and spring winds influences the morphology by producing:

- secondary and tertiary slipfaces;
- rearward crestal migration and therefore separation of crest and brink;
- non-equilibrated convexity of form;
- and eastern horn elongation where the 2D planimetric shape of the dunes becomes asymmetrical.

The interplay of variable wind directions and intensities, in accordance with any slight variation in slope gradient leads to changes in the morpho-dynamic behaviour of the dunes, which is ultimately expressed in dune morphology and sedimentology. Subtle changes in the slope gradient influence feedback adjustments between dune morphology and wind regime, causing a continuum of variation in the shear stress and the power available for sand transport, erosion and deposition. Both the morphology and sedimentology of the Gurra Gurra crescentic dunes undergo perpetual change. Furthermore, alteration to either wind intensity or wind direction individually will produce different morphologies for any sand sea. Change to both, will develop interacting complexities that can only be solved through repeated seasonal observations and an awareness of the continuum of spatial and temporal variation that occurs in all global dunefields.

7.5 Substrate Cohesiveness and Pavement Structure

The nature of the physiographic setting and its control in the development of dune morphology has received comment from Norris (1966), Theileg (1978a; 1978b), Mabbut and Wooding (1983) and Lancaster (1988a). Norris (1966) has suggested that 'subtle topographic influences' may control dune morphology, while Mabbutt and Wooding (1983, p.68) commented that substrate changes "account for fairly rapid adjustments in equilibrium dune pattern". Theileg (1978a) made similar inferences, suggesting that the smoothness of the bedrock is one of several phenomena contributing to the limited extent of the dunes in the Salton Sand Sea. Having already demonstrated the morphological variations of the

Gurra Gurra dunes to be due to seasonal changes in the aeolian regime, the following section discusses the changes influenced by seasonal modification of the surface texture between dune and intradune areas.

Surface texture or roughness is principally controlled by grain-size, soil-regolith moisture content, geochemical bonding agents such as clays, salts or organic matter, the presence of surface crusts, as well as the vegetation type and areal population. As a result of these variables, the degree of surface roughness can alter, which in turn modifies turbulence, the height of the saltation layer and the saturation level of the saltation curtain. Major mechanisms of change to surface roughness at Gurra Gurra waterhole involve the inter-seasonal contrasts of:

- vegetation
- coarse sand lag deposits
- montmorillinite and kaolinite content of the sand
- moisture content of the dune sand

Of these four parameters, the last two are most influential in the genesis and morphologic evolution of the Gurra Gurra crescentic dunes. The mechanism of interparticle bonding and substrate cohesiveness was found to be a simple condition of clay content and seasonal variation in the wetting-drying process from which more erosion-resistant surface crusts could form. Variable surface roughness is in part, due to seasonal textural contrasts between the uniform structure of the undesiccated but moist, clay intradune floor in winter, and the heavily desiccated dry, clay-sandy intradune in late spring-cum-summer. The seasonal alternation of intradune and aerodynamic roughness is a primary factor in establishing and maintaining morphological equilibrium. It is surface roughness and, hence, shear stress per unit area of the wind that causes changes in turbulence, flow acceleration or retardation, saltation height and the density of the saltation curtain. In other words, surface texture effects a control on sand mobility. Therefore, adjustment of the mobile crescentic dune form to changing wind regimes fundamentally involves the contrast of texture between the dune and intradune.

The moisture content of the sands can also be inhibiting or conducive to the growth of shrubs and ground hugging plants[1], which in themselves alter the surface roughness of the intradune, and behave as air-retarding obstructions to wind flow. The dunes of autumn clearly demonstrated the interaction of dune morphology and vegetation. As an obstacle to air flow, both living and dead shrubs caused the southern dunes to develop a poor, less distinct crescentic form. In contrast, the northern transverse dunes maintained a morphological structure that was little influenced by the relatively sparse vegetation (also refer to Chapter 7, Section 7.0.2).

In support of these observations is the report of Mabbut and Wooding (1983) who described crescentic dunes developed upon the floodplain of the Finke River in the north-western Simpson Desert of the Northern Territory, Australia. The physiographic setting for the Strzelecki crescentic dunes not being too dissimilar to the 'transverse trains' and 'network' dunes at sites along the Finke River. At both sites the crescentic dunes occur within close proximity to a major stream channel and or flood-plain. However, unlike the Finke River crescentic dunes, the Strzelecki transverse forms do not derive significant amounts of sediment from the neighbouring alluvial source, nor do they continue across and beyond the channel. Nonetheless, both localities demonstrate small but relatively highly populated sites of crescentic dunes.

In both settings, the common physiographic parameter is the unconsolidated substrate of thick sand upon which the crescentic dunes develop. The thickness of the sand profile is important at both sites in that if the textural contrast is to remain static over a considerable dune building episode, an abundant deposit of sand is required. That is, the aerodynamic

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[1] It is important to recognise that the immobility of Australian desert dunes is a consequence of low windiness, rather than stabilisation by vegetation (Ash and Wasson, 1983).

roughness length, and hence surface roughness must be preserved for an adequate distance. At Finke both linear and crescentic morphologies co-exist, and develop under the same regional aeolian regime. The evolution of the two morphologies seems to be equally as related to substrate texture and the annual directions of sand drift potential, as only the former parameter significantly alters for the site. Similarly, the Strzelecki Creek site exhibits dune building into morphologically diverse barchanoid, transverse and short segment seif-linear landforms that can be interpreted as responses to variable wind flow across the relatively unconsolidated underlying linear dune sands, from which the crescentic forms arise.

It is presumed that once a juvenile crescentic form has been initiated upon the clay-enriched and smoother intradune, it is the seasonal variation in the substrate cohesiveness that assists in controlling the dimensions of these dunes under the influence of multi-directional winds. In this manner, both the Finke River and Strzelecki Creek crescentic dunes may have similar origins. The Finke dunes evolve to linear dunes of increasing complexity (see Chapter 2, Section 2.7.3) as the substrate consolidates into alluvial gravels or gibber plain. In many respects the evolution of the Finke River linear dune morphologies fits the mechanism proposed by Kar (1987) in the Thar Desert and the evolution of fulji as described by Bagnold (1965). In these cases, it is reported that crescentic dunes interlink, toe-to-stoss, forming hollows (fulji) that later infill to form longitudinal ridges.

The morphology of dunes built upon consolidated substrates and in near uni-directional winds, for example, the Salton Sand Sea and the Pampa de la Joya barchans, do not occur in a similar morphodynamic setting as that of Strzelecki Creek. For this reason there are few morphometric similarities between these sites. Crescentic dunes in dimensional equilibrium (i.e. height-to-width-to-length ratio = 0.1:1:1.5), develop in areas of sufficient, but not copious, sand supply and upon desert pavements that are relatively consolidated compared with the dune's surface texture (also refer to Chapter 7, Section 7.4). 'Hard' pavements assist in maintaining equilibrium of form by offering least resistance to sand mobility, facilitating the adjustment of form due to the aerodynamic roughness offering less flow turbulence and resistance. Moreover, saltation energies are not decelerated as significantly as for less cohesive 'soft' surfaces, nor are grain impact energies dissipated as much as on less fluid substrates. Thus, there is less deterioration to saltation frequencies, heights and saturation levels of the saltation curtain.

In the light of the above factors, a crescentic dune will only establish morpho-dynamic equilibrium:

- if there is an insignificant annual variation in wind direction;
- if the dune-substrate surface roughness is significantly different;
- and if substrate roughness remains relatively homogenous, thereby maintaining aerodynamic roughness for the length of the surface.

Under the influence of multi-directional winds, a quasi-equilibrated form occurs only if the interdune physiography and surface roughness allows rapid dimensional adjustments of the dune mass. Where the interdune is of similar textural contrast to the dune body, rapid adjustments to the dune mass are less likely, due to the ease of erosion and the mobile nature of the substrate. As such, merging of the dune and substrate leads to the construction of amorphous, transient structures. Such an environment exists on the flanks of the juvenile dunefield of Gurra Gurra (Fig. 4.2), where amorphous sand masses with transient slipfaces occur upon a deep sedimentary pile of unconsolidated sand. Only where a cohesive intradune floor exists, composed of moistened clay-enriched desiccation polygons, do individual and distinct crescentic dunes develop. Since the mid-1980s when the image used to produce Fig. 4.2 was photographed, dune-field organisation has become more obvious, with increased downcutting of the parent linear dune and, hence, the development of a wider clay-enhanced intradune floor and less amorphous sand masses. This has caused dune growth and, specifically dune organisation, into transverse ridges at the northern-most end of the field. The maintenance of equilibrium morphology is most difficult in an environment where there is high variability in wind directions and, wind intensities, and where the dune and interdune textures lack contrast. The existence of the

Gurra Gurra crescentic dunes is, therefore, unique and highlights the site specific environmental and physiographic setting in which these crescentic dunes develop.

7.6 Desiccation-Polygon Erosion and Yardang Evolution

The development of sustainable crescentic dunes upon an individual linear dune in the Strzelecki Desert is correlated with the dimensions of the underlying linear dune, its geometry and associated upslope velocity amplification inherent to these parameters. Notwithstanding these observations, the presence of clay-rich horizons within the indurated floor, upon which seasonally-controlled desiccation polygons occur, are vital to the genesis of crescentic forms across the breadth of the linear landform.

Desiccation polygons occur on the surface of the intradune floor and result from dewatering the montmorillinite-kaolinite content of the sands under the influence of spring and summer temperatures. As suggested by Chico (1963) the type of clay and its cohesive properties influence the extent of contraction. Montmorillinite-rich sediments are able to contract more than kaolinitic sands. Tensile stresses of the intradune area develop a regular fracture network that is non-orthogonal, hexagonal and complete in type (Allen, 1982a). The general profile of the polygon is very slightly concave-up and approximately 50 cm wide. Secondary polygons, internal to the primary form, measure approximately 10 cm in diameter and produce poorly developed mud-curls, although at the toe of several dunes in autumn, these clay-rich laminae show well-developed convex-up drying and curling. These features reveal the original source of the dune sands to be muddy alluvium, a finding also reached by Wasson (1983a). This is typical of the linear dunes west of Strzelecki Creek. According to Allen (1982a, p. 550), the genesis of desiccation features is a function of the stresses created by surface tension on the films of water that bridge mineral grains at bulk water-contents lower than the sediment dry porosity. The stresses are at a maximum with a water-content intermediate between full saturation when the forces are zero, and complete dryness, also when the forces are zero.

The development of hexagonal polygons upon the horizontal, but continuously downcutting intradune area, illustrate that the maximum release of strain energy per unit area has been achieved. It implies that fissuring commences at many points simultaneously with the subsequent rapid development of fractures. This is a phenomenon not typical of most shrinkage cracks (Allen, 1982a). The lateral extent of the polygons and the uniformity of their dimensions, together indicate a clay layer of relatively uniform thickness and a homogenous and brittle sediment that dewaters at a common spatial rate Lachenbruch, 1962; Allen, 1982a) (see Fig. 5.7).

If it is assumed that polygon width is within the generally accepted range of 3-8 times the crack depth or layer thickness (Allen, 1982a), then a depth of between 6-17 cm should be expected here. Although some infilling of the fractures had occurred, such depths have been observed and are seen as ideal venturi conduits which can channel and increase surface friction velocities. Erosion of the fracture is assisted not so much by laminar flow through the fracture, but by the development of turbulent flow at the boundaries (boundary-layer turbulence of the fractures), as well as by secondary flow or horseshoe vortices being produced by flow redirection upon contact with the face or apex of a polygon. Work by Greeley and Iversen (1987) have shown that the strongest vortices shed from the upper surface of an obstacle occur when the wind vector bisects the surface edge of the obstruction. Shear stresses associated with such vortices are extremely high. The power of such forces is amply demonstrated by the extent of wind scour in loose sediments around nebkha in summer. The combination of surface vortices, shed by the bisection of the primary summer winds, combined with horseshoe vortices channelled either side of the obstacle, and assisted by venturi flow, imply that complex flow patterns aid the rapid downcutting and undercutting of desiccation fractures. Whitney (1983) offers support for this argument, having identified that moats and channelled flow around the mega-yardangs of the Kharga Depression in Egypt, develop venturi flow that accelerates erosion along these paths.

Once a significant vertical profile of the polygon has been established within the boundary layer, flow separation and reverse flow may also occur at the air-obstacle interface upon the windward face, the upper surface of the polygon and at the rear or leeward slope. Polygon structure is such that linear and orthogonal faces develop between the top plane and the lowering intradune floor. As a greater surface area of the polygonal faces is exposed, fallen, unconsolidated deposits are removed, thereby maintaining the morphology of the hexagon. This would form a typical pedestal shape that correlates with the development of the closely spaced yardang pedestals at Gurra Gurra waterhole. Ward and Greeley (1984), using wind tunnel and field studies, identified that abrasion was greatest upon the windward slope, initially cutting at the corners and then sequentially eroding the windward slope itself. The hexagonal form and sharp boundaries of a desiccation polygon's frontal sides offer an ideal obstacle which would fit the windward abrasion mechanism suggested by these authors.

Repeatedly, yardangs were seen to have a surface capping of clay laminae, a possible relict of the original 'parent polygon', while yardang relicts align themselves equidistantly behind each other, which may be representative of distances between the centres of individual polygons (see Fig. 5.5 & 5.6). The heights of yardang groups were also very similar, indicating a common plane from which erosion commenced and demonstrated the several cm a^{-1} rate of downcutting observed (also refer to Chapter 5, Section 5.3).

The typical width-to-length ratio of 1:10 for the Gurra Gurra yardangs is due to the interrelationships between polygon dimensions and form, differential erosion between the relatively low cohesiveness of the dune sand and the cohesive nature of the clay intradune layer(s), as well as the azimuth and gradient of the linear dune's internal cross-bedding. This scenario is one which accounts for the genesis, form, spacing and dimensions of the yardangs. Similarly, Ward and Greeley (1984) and Halimov and Frezer (1989) identified yardangs with the same dimensions at the Rogers Lake playa, California and in the Qaidam Depression of central Asia, respectively. Other streamlined 'islands' such as those formed by fluvial erosion (Baker, 1978a, 1978b, 1979) also demonstrate a particular range of planimetric ratios. The genesis and evolution of yardangs and 'streamlined islands' thus involve similar processes. Uni-directional currents, cohesive materials and channelling effects are characteristic of such landforms, regardless of size, geographical location and lithology. The only different variable affecting yardang geometry, is that of time and the rates of erosion for specific localities. As for crescentic dune-forms, the environmental setting is fundamental to attaining and sustaining equilibrium morphology of yardangs.

8 Granulometric Differentiation on Crescentic Dunes

8.1 Sorting Processes

This section identifies the processes here considered to be responsible for surficial sedimentological changes at the meso-dune scale. The differentiation or sorting of dune sediment and its expression as a process-specific, surficial-sedimentological signature, involves the morphological and mineralogical characteristics of the sediments, the aeolian regime and the micro-morphology of the dune. The preceding chapters have identified the nature of the seasonal wind patterns, dune morphology and prevalent quartzose mineralogy of the sediments. From these, it is possible to demonstrate micro-morphologic zones of erosion and deposition on the dunes.

Processes of sediment differentiation observed for the crescentic dunes of Gurra Gurra are:

- winnowing;
- sifting;
- interstitial entrapment and protection;
- size segregation;
- avalanching;
- slope threshold sorting.

All six processes operate, though not necessarily simultaneously, and involve the entire dune when high energy summer winds are active, or only the upper crest-brink region when u_* is relatively low, as for winter. Figure 4.1 in Chapter 4, illustrates the cyclical morphologies of the Gurra Gurra dunes within a variable reversing aeolian regime, and demonstrates the transitions of form that involve simultaneous sedimentological transformations. Morphology assists the development of the sedimentary distribution in that changes in slope gradient influence the operation and type of process.

Differences between sifting and size segregation are related to the energy thresholds of the aeolian environment. Size segregation is achieved by the impact of saltating grains and the transfer of impact energies to larger underlying grains, followed by compressional rebound of the subsurface grains to the surface. Sifting only occurs, however, when the fine grains settle amongst the interstices of larger surface grains. Suspension loads that do not impart significant impact energies and granular displacement of the surface laminae are typical of this process. It is most probable that both processes often operate in tandem; finer grain-sizes are vertically sorted with the upward movement of larger grains. The process of fluid threshold motion may also be of some importance in the sifting process. Iversen and White (1982) found that as the velocity increases to the point of threshold motion, particles begin to vibrate with increasing intensity, until a critical frequency initiates departure from the surface. Such vibrations may be significant in assisting the interstitial sorting of the surface laminae. Winnowing and interstitial entrapment are major sorting processes on the dunes of Gurra Gurra. They reflect the ability of the wind to remove or restrict smaller grains from between the intergranular voids of larger resting grains of the uppermost laminae. Removal of the grains is a complex process that involves:

- grain-size;
- roundness and angularity;
- density;
- interparticle cohesiveness due to moisture and electrostatic forces between grains;
- wind velocity;
- slope angle of the surface, and,
- the degree of projection-exposure into the ambient air flow at the boundary layer.

Winnowing, sifting, and interstitial entrapment are major mechanisms involved in the cyclical transport of fines on the dunes of Gurra Gurra. These sorting processes predominantly involve the redistribution of fines to-and-from the crest-brink region, even when seasonally low intensity conditions do not favour the processes of avalanching, size segregation and slope sorting. Through these processes, the third moment or fine tail of the distribution, which is most sensitive to small changes in grain-size distribution, infers details of the transport process. Friedman and Sanders (1978) and Pettijohn *et al.* (1987) offer support for such reasoning (see Chapter 3, Section 3.5.2.3). Overall, very few dune samples demonstrate bimodal distributions, as compared to the strongly bimodal and polymodal natures of the reg (interdune) and fluvial sediments respectively (see Chapter 6, Section 6.0). The positively skewed distributions, however, depict the existence of a volumetrically small tail of fines amongst the predominantly coarser grains. Such a tail of fines may be attributed to deposition of a suspension load mantling the dunes and settling in the inter-particle voids of the coarser sand. Warren (1971) explained the bimodal nature of the sands of the Tenéré Desert of NE Niger, as due to the saltation and removal of intermediate sand sizes that do not settle amongst the creep fractions. This encouraged the finer fractions in saltation and suspension to settle between the inter-particle voids of the creep load, thereby protecting them from further entrainment.

8.1.1 Upper Dune

Irrefutably, the sedimentology of the Gurra Gurra dunes reflects that throughout a year, greatest activity occurs on the upper dune elements. Excepting for summer and spring, the lower dune elements are relatively inactive due to the low windiness for these seasons. A significant u_* is achieved upon the higher elements, produced by the convergence of upslope air flow. In accordance with this, Lancaster (1985a) also found that the rate of sand transport is significantly less at the windward base compared to that on the crest; light winds can lower the crest, forming a separated crest-brink, while the windward toe remains inactive. Similarly, it can be inferred that under reversed wind conditions at Gurra Gurra, the crestal zone remains under active shear, whereas the base of the leeward slope is inactive.

During the metamorphosis of the sharp accentuated shape of the summer dunes to the broad convex dune forms of winter, there is a cycle of erosion and transport of fine-intermediate sand from the coincident crest-brink zone, with contemporaneous deposition onto the stoss slope; the process of preferential sorting. A plausible mechanism for this type of distribution, is offered by the preferential winnowing from a crest-brink of an intermediate grain-size onto a windward slope. This leaves behind an interstitial fraction of very fine sands amongst a predominantly coarser grain load that is not capable of being lifted by the available threshold friction speed of the wind. This observation coincides with that of Bagnold and Barndorff-Nielson (1980), who noted that deflation of the smaller grains develops an erosional surface that is overloaded with coarse grains. As a consequence, crestal sediment is distinctively well-sorted but fine-skewed.

Additionally, sorting upon the upper stoss slope is achieved by reverse tangential avalanching, grading down the rear slipface, with possible minor but very localised sorting by a small rearward reverse eddy that winnows fines upslope towards the base of the rear slipface. Although the reversed winds are generally of low intensity, the acute angle (26°) and height (*c.* 1m) of the rear slipface may cause separation and reverse flow. This may be especially characteristic of strong storm winds. Most likely the mid-winter crests portray a sorting process midway between developing a bimodal distribution as for the Tenéré sands (Warren, 1971), or that due to incomplete winnowing, a coarse grained very well-sorted, negative or negligible skewness deposit will develop. The aeolian cycle is not a compartmentalised process, but represents a gradual change initiated by seasonal wind regimes. The signature of sorting by the end of winter may be identified as an end member distribution, but may remain an intermediate one, if the aeolian regime alters little throughout the winter. As Watson (1986) has explained, grain-size variation is attributable to the wind regime at the time of sampling, and deflation can leave a residue of coarser sand.

The coincident and sharply accentuated linear lee and stoss morphology of the summer dunes, and to a lesser extent the autumn dunes, are developed under high frequency and high intensity winds. Strong summer winds erode the windward slope, redistributing a turbulent mixture of grain-sizes over the upper stoss and crestal elements. This produces an upper dune that is intermediate in grain-size and sorting. Subsequently, the removal of fines from the stoss flank leaves a coarse, poorly sorted windward element (see Chapter 6, Section 6.6.3). The high impact saltation energies over the entire dune causes granulometric transitions due to grain-size segregation on the stoss and crest-brink, and grainfall and avalanching on the lee. However, each of these processes are in constant flux due to the peak of dune mobility.

There is no statistical difference in crestal sorting on either the lee or stoss slopes. Crestal sorting in summer, however, represents an intermediate value between the statistically significant result of a poorer-sorted stoss and a better-sorted lower lee. Under the higher threshold velocities of the southerly summer winds, the vigorous inter-mixed nature of the creep, saltation and suspension loads produces the more poorly sorted, but negligibly skewed distributions of the windward to crestal dune positions. Skewness is not significantly altered over the dune except on the eastern horn, where the sediments are less skewed than on the rest of the dune. This further identifies the chaotic activity of the entire dune in summer.

In mid-to-late spring the morphology of the summer dunes begins to reform as the winds turn south-easterly and transport the finer surficial sands of the winter stoss slope back onto the crestal zone. Prior to the onslaught of intense spring and summer winds, when granulometric inter-mixing is rife, deposition upon the crest may be influenced by the increased surface roughness of the inherited coarser winter surface. Consequently, fining and smoothing of the crest may occur. Changes in surface roughness and changes in shear velocity on meso-scale terrains have been documented by Greeley and Iversen (1985, 1986) for the Amboy lava field in the Mojave Desert. Here it was found that the leading edge of a discontinuous change between smoother alluvial and rougher basaltic surfaces, u_* temporarily averages 44 percent higher, with a zone of erosion predominating. It is uncertain, however, whether the effects of a changing surface roughness occur at the inter-granular scale. Cooke *et al.* (1993, p. 249), in discussing the relationship of the steepness of the curve of threshold shear velocity to the decreasing size of fine particles, suggested that the effects of surface roughness are not important at the granular scale. Nevertheless, as summer approaches, erosion of the transitory rear slipface and development of a lee separation bubble lead to the eventual re-development of the principal advancing slipface at the angle of repose (32°), the closure of the separated crest-brink and the forward migration of the crest with a subsequent return to morpho-dynamic equilibrium.

8.1.2 Lee Slope

The sedimentological character of the lee slope in winter is characterised by upslope granulometric differentiation under the influence of high threshold shear over the upper lee (brink), as indicated by very coarse sands, a high proportion of sorting and high positive skewness. Both preferential winnowing of intermediate fines and the protection of diminutive fines by inter-particle voids operate to varying efficiencies under the different seasonal aeolian patterns. The finer, relatively less sorted and very low positive skewness of the mid to lower lee slope may be partly attributed to inheritance from the summer grain-size distributions. If aeolian activity is quiescent throughout autumn, the summer sedimentology due to grainfall and avalanching, may be partially preserved due to the absence of significant winter u_* on the lower morphologic locations.

Conversely, during both autumn and spring there is downslope coarsening and a lessened skewness for the lower lee. This sedimentological character correlates with significant oblique wind components across the lee face whereby fines are either winnowed out, leaving behind a coarser erosion lag surface, or diffused to the underlying laminae by avalanche grading. In spring, widespread avalanching takes the form of many narrow sand-flow tongues, resembling a 'bottle-neck' shape, that originate and gouge and notch the crestline. These tongues subsequently descend the slipface along a narrow front, characteristically with a convex upward surface that rides above the slope. The dominant presence of avalanche tongues on the lee and rear slipface in spring and rear slipface in winter, indicates active surfaces from which reverse tangential grading develops. Reverse tangential grading results in coarser grain-sizes being deposited upon lower levels of the slipface, and is apparently due to larger grains being displaced towards the free surface of the flow by reverse perpendicular displacement during the flow (Middleton, 1970; Allen, 1982b). In other words, the smaller grains filter down diffusely between the larger grains due to both strain and vibration dilating the grain mass during flow. The relatively higher velocity of the larger grains, as well as flow over a bed of finer grains with a lower surface roughness, assists in transportation of the larger grain-sizes downslope. Reverse tangential grading produces the sedimentological signature of coarser, but poorly sorted and less finely skewed granulometric distributions downslope, as observed in spring. It is also noteworthy that the lower inclination of the lee slope in both winter and spring is also due to the process of avalanche.

The bi-modal action of the diurnal spring winds induces a complex and poorer lower dune sorting, than for those seasons with only one significant wind orientation. Decreasing lee slope sorting and lower fine skewness occur during reversing winds in both winter and spring, with evidence of upslope-to-downslope slumping of the upper lee by avalanching. In spring there are also greater shear and motion of the lee slope by the activity of strong oblique lee-cutting winds, as previously discussed. Poorer down-slope sorting and declining skewness is most likely due to initial grainfall of coarser sands, exposed and mobilised when the ambient reversed and oblique winds winnow and subsequently de-stabilise the upper zone of the principal slipface and alter the amount of crest-brink separation. Although the wind direction is diametrically opposed to the principal direction of advance, grainfall down the lee slope still persists under the more quiescent conditions of winter, due to the slipface angle remaining significantly high (max. 29°) and the actions of disruptive intermittent storm gusts. Destabilisation of the lee slope is demonstrated by minor 'bottle-neck' sandflows, cohesive raft-like slumps and ripple migration over the upper dune as observed during the intense reversed winds of a winter thunderstorm. This also occurs in spring when there is oblique sand movement up and across the entire slipface in the afternoons. In like manner, Greeley and Iversen (1987) reported reversed transport by the leeward eddy upon the slipface in the Kelso dunes, California.

Further evidence of lee slope avalanching is revealed by the presence of accumulated sand aprons at the base of the barchanoid segments. These aprons are not present in summer or autumn when the crest-brink morphology is coincident, although air flow and consequentially grain-fall over the crest and onto the lee slope is prevalent in these seasons. This indicates that the apron pile is a deposit which is not recognisably associated with the aeolian regime of dune advance in late spring and early-to-mid summer. Wind shear is at a minimum at the base of the lee slope during these periods. Consequentially, the avalanche apron deposits are characteristic of only reversed and oblique aeolian wind patterns, when the entire leeward slope is under persistent assault by the wind.

8.1.3 Stoss Flank

The winter upper stoss accumulates the fines winnowed from the crest. Similarly, the base of the slope demonstrates an association with very fine sediments that could have originated from the upper dune. Kocurek *et al.,* (1990) observed slopes as low as 10° undergoing grain-fall. The Strzelecki winter stoss slope, in part, typically has angles ≥ 10°, which may assist in fining of sediments along its length in reversed winds. Better sorting is associated with the activity of the upper dune, once again identifying erosive differentiation, with the transport and intermixing of various size grades onto the lower stoss. The existence of a weak separation eddy would assist in the turbulent mixing of the finer fractions over the mid-to-upper stoss flank. The less positively skewed character of the toe of the windward slope identifies the cessation of downslope entrainment of fines prior to reaching this area, and the inactivity of aeolian processes on the toe at this time. However, when the winds are reversed, a veneer of fines eroded from the crest and re-deposited upon the stoss, are sufficient to alter the surficial sedimentology of the coarser basal creep fractions.

In contrast to the winter stoss, the summer and early autumn stoss is coarse and poorly sorted under ambient high winds. The degree of sorting on the windward slope reflects the turbulent character of transport under the intense southerly summer winds, and is contrary to the influence of reversed winter winds. Improved sorting between the lower stoss and crest-brink and lee components of summer, suggests more uniform sorting over the entire near-equilibrated dune. Both the summer and autumn distributions reflect this process, with the late summer signatures being preserved on the early autumn dunes. Similar observations have also been reported by Lancaster (1981c; 1982b; 1989) for both the coastal and inland dunes of the Namib sand sea. Skewness for the dunes of Gurra Gurra is greatest on the lower stoss of the summer dunes, while all dunes in winter, autumn and spring reveal less positive skewness. The presence of such low positive skewness is considered pertinent to the low level of activity upon the lower dune in these seasons, contrary to the more vigorous activity of summer. The action of dune migration and mass movement in summer, is signified by the removal of particles from the stoss, which readily saltate. They leave behind both the larger and the finest particles which are afforded inter-granular protection. However, as the distributions remain significantly unimodal rather than bimodal, a range of grain-sizes must be steadily transported, leaving behind a low weight mass of fines overall. The lessened mobility of the sediment in all other seasons, is therefore interpreted as illustrating a broader range of grain-sizes inhabiting the lower stoss, and aptly demonstrates the contrast between crestal activity and lower dune inactivity throughout most of the year.

8.1.4 Dune Horns

The horns of the winter, summer, autumn and spring dunes show no statistically significant distinction in mean grain-size between micro-morphologic locations, except for the comparison of the low lee and low horn-east in autumn. It is suspected that this one significant spatial variation is a consequence of the winnowing of fines from the horn under the influence of seasonally nascent oblique and horn elongating winds.

Sorting during winter is better on both the eastern horn and crest relative to the stoss, western horn and lee, while the western horn is intermediate between the best and worst sorted locations. The difference between the lee and eastern horn is statistically decisive.

Similarly, sorting upon the eastern horn is high in summer and spring. The higher degree of sorting on the eastern horn is correlated with the elongation of the eastern arm. The action of prevalent westerly oblique winds assists in stunting the western horn, while pronounced sedimentological differentiation occurs along the increasing length of the elongating eastern horn. Accordingly, winnowing of finer material from the eastern horn develops a better sorted sediment sample and the protection of diminutive fines between inter-granular spaces would demonstrate a well developed positive skewness. Such skew values are pronounced for the eastern horn in all seasons. The balance or contrast between the degree of sorting and the degree of skewness is related to the quantity of winnowing relative to intergranular protection.

8.2 Concluding Remarks

The sedimentological characters of the Gurra Gurra dunes is one of subtle seasonal contrasts, that can be attributed to either the removal or accumulation of the fine tail of the sedimentary grain-size distributions. Seasonal variation is gradual and transitional. For example, the sedimentary distributions of the early-autumn dunes maintain the summer character and are illustrative of continuity of the same processes. The coincident crest-brink for both seasons demonstrates the finest relative mean grain-size distributions. Autumn is the least active season when minimum saltation occurs. Hence, the fine crests and relatively coarse stoss slopes must depict the characteristics of late summer, when the winds are of high frequency and intensity. It is concluded that the samples from the autumn dunes reflect the inherited features of late summer, and that opposite distributions feature at the end of autumn-early-winter.

The presence of juvenile morphologic changes in the mid-autumn observations, indicates that more subtle and far less intense winds were beginning to reverse, but were not significant enough to alter the gross morphology or the sedimentology inherited from the previous season. These morphologic changes and the absence of sufficiently strong winds provide strong evidence of the autumn sedimentology being an inherited signature rather than being idiosyncratic of that season. By late autumn, the morphology and sedimentology begin to reflect a typical winter distribution, that fully develops by late winter and is sustained throughout early spring. Subsequently, the late winter temporal and spatial attributes begin to alter by late spring, when the summer signatures start to develop. Again the process of seasonal variation is emphasised as being gradational and not compartmentalised into discrete temporal occurrences. These gradual processes of change are in themselves variable and dependent upon the annual weather patterns.

Chapter 6 has shown that the variations of surficial sedimentology between like micro-geomorphologic elements in different seasons, are less than the variation between different micro-geomorphologic locations in the same season. As emphasised, summer and late spring are times of erosion and deposition over the entire dune, while winter and autumn dunes undergo sedimentologic changes to varying extents, that are restricted mostly to the upper dune areas where low intensity winds are amplified to threshold shear velocity by upslope convergence of the flow lines. Even then, amplified shear velocities are low and often intermittent, and restricted in their saltation capacity. Such wind energies are capable of removing fine-to-intermediate sands by winnowing the crestal zone and transporting them to other locations on the dune. This leaves behind a residual surface layer of coarser grains mixed with very fine inter-granular sands. The deposition or protection of existing very fine materials amongst the inter-particle voids of coarse grains, contemporaneously sifts or size segregates these materials. Such processes are encompassed within the 'protectionist theory'.

In further support of this scenario is the lack of gross temporal sedimentological variation, as well as the significant differences in crestal configuration for each season. There is no significant variation between the inter-seasonal contrast of mean grain-size and skewness for the crest-brink and low lee, and for skewness on the low stoss. There has therefore been no drastic alteration in the fundamental dune sedimentology since dune formation. Significant changes in the mean grain-size of the lower stoss are explained by differences caused by the accumulation of coarse (≤ 1.0 ϕ) intra-dune lags, most probably of fluvial origin, upon the lower stoss during the high intensity winds of early-to-mid summer and late spring. In contrast, the removal of all fractions upon the lower dune, excepting for those that are only capable of creep, develop distributions typical of the dune fringe in summer and spring. It is commonly reported that the coarsest grains dominate the fringe of the dune rather than the upslope elements (Finkel, 1959; Hastenrath, 1967; Warren, 1976; Lancaster, 1982b; Watson, 1986).

Nevertheless, the time of sampling and the ease with which fines are eroded and deposited, can drastically alter the surficial grain-size distribution for a given locality. Accordingly, the inter-mixing of fines with the lag deposits in winter and autumn, mask the granulometric impact of the coarser fractions in the other two windier seasons. Figures 6.1 to 6.3 in Chapter 6 adequately illustrate the higher weight percent of coarser (≤ 2.0 ϕ) particles in the stoss distributions relative to the percentage of the same fractions in the upper dune.

Contrary to the lack of temporal differences for the first and third moments are the many significant inter-seasonal contrasts for the second moment (sorting) (see Chapter 6, Section 6.3). This result clearly supports the idea of the relocation of intermediate-fine fractions between elements of the dune on a seasonally cyclic basis, which can be effected by an active upper crestal region relative to less active dune positions for most of the annual cycle. Such a mechanism alters little the general formative sedimentology of the dune, but rather recycles size specific fractions within the existing dune system. Gross sedimentological change through transfer from outside of the dune system is restricted to the aforementioned input of dune creep load. Thus, combined with specific sorting mechanisms and upper dune changes in crestal morphology, the seasonal recycling and transfer of intermediate-fines between the upper and lower micro-morphologic locations, provides a model that explains the spatial and temporal sedimentological variability of crescentic dunes such as those at Gurra Gurra waterhole.

9 Conclusion

The research objectives of this thesis were designed to explore the spatial and temporal changes and interactions of both dune morphology and dune surficial sedimentology for the crescentic dunes of Gurra Gurra waterhole in the Strzelecki Desert of South Australia. Specifically, the focus was to examine the spatial variation of dune morphology using dimensional differences compared to crescentic dunes in other terrestrial sand seas and the dunefields of Mars, and the comparative micro-geomorphic changes of element morphology in different seasons. Other objectives were to compare dissimilar micro-geomorphological elements and their surficial sedimentological distributions in each of the four seasons, and the surficial sedimentology of similar micro-geomorphic elements in all seasons. These objectives were achieved.

The examination of the spatial and temporal comparisons between dune morphology and sedimentology has lead to a conceptual model of genesis and evolution for the crescentic dunescape of Gurra Gurra waterhole, as well as a discussion of similar processes being active in other terrestrial and extra-terrestrial crescentic dune-fields.

Research in many desert regions has shown that the genesis and evolution of crescentic dunes are responses to the fundamental parameters of:

- a reliable sediment source;
- a variable surface roughness, due to inter-seasonal contrasts of temperature, moisture and vegetation;
- a primary uni-directional regional aeolian regime.

In broad view, these factors are also responsible for the form and development of the Gurra Gurra dunes.

In detail though, the intricate natures of dune micro-morphology and surficial sedimentology on the Gurra Gurra dunes are responses to site specific sand sources in a semi-closed geomorphic system, topographic enhancement and seasonal periodicity of less dominant secondary and tertiary wind directions, strengths and durations, as well as other climatic and vegetative influences.

9.1 Morphology and Morphometry

Figure 9.1 demonstrates the pathways between the aforementioned variables that are considered fundamental in the genesis, macro-morphologic evolution, cyclical micro-morphologic and surficial sedimentologic development of the Gurra Gurra crescentic dunes. The greater dimensions of the underlying linear source dune, relative to others in the region, enhance oblique upslope wind speed amplification over the eastern plinth of this dune in summer. This, in turn, increases the ability of the wind to lower the linear form. Downcutting is evidenced by the lowering of the intradune floor and the formation of yardangs due to venturi flow along polygonal desiccation fractures of the clay-enriched surface. Desiccation caused by the swelling and shrinking of eluviated montmorillinitic laminae amongst the quartz rich sands, is a primary process providing a sediment source. The sediment released is redistributed downwind into crescentic dunes that are also developing upon sites of variable surface roughness. Such processes have lead to significant lowering of the northern relief of the linear dune and to the development of stacked transverse dunes.

The aforementioned parameters are also involved in shaping the micro-morphology and surficial sedimentology of individual crescentic dunes. Regional airflow over the linear dune is further intensified by the influence of the evolving crescentic dunes, while variation in local (intradune) surface roughness acts not only as an initiation site for the genesis of dunes, but also affects the rate at which dunes advance and maintain form, and the quantity of sediment transferred into the dunes. Surface roughness is also a consequence of the type, size and distribution of vegetation upon the intradune area. Specifically, micro-morpho-sedimentary change on the crescentic dunes is a product of:

- the multi-directional and multi-intensity seasonal aeolian regime;

- upslope wind speed amplification over either the lee or stoss slopes depending on the seasonal wind direction.

Such influences control the development of morphological features such as:

- rear slipfaces;
- the degree of convexity and concavity of the lee-to-crest and stoss-to-crest slopes;
- the sequential separation and coincidence of crest and brink elements;
- horn elongation or diminution;
- dune width, length and height;
- slipface and stoss inclination;
- apron deposits;
- lee projections.

Similarly, sedimentological change is also affected by micro-meteorological conditions and is evidenced by the degree of winnowing of intermediate grain-sizes and the entrapment of fines, especially upon the upper dune, and the distribution of creep load around the perimeter and lower slopes of the dunes and intradune areas. Further to this, the orientations of yardangs, scour flutes and ripple patterns attest to the specificity of the micro-meteorological conditions at Gurra Gurra waterhole.

Through both the influence of multi-directional winds and the amplification of shear velocities, both dune form and sedimentology are in continual transition between the dynamic end-members of equilibrium and quasi-equilibrium. Such dynamic antithesis is depicted in the generally poor mathematical correlations between different morphometric parameters. Simple bivariate relationships between dune length, width, wavelength and height are nebulous at best. The elongation of the eastern horn and subsequent evolution of minor seif-linear type dunes distort dune width, which is a factor that assists in the dimensional impropriety of crescentic form. Similarly, comparative dimensional analysis with the dunes of Pampa de la Joya and the Salton Sand Sea show negligible correlation. The combined action of multi-directional, variable intensity winds and the textural contrasts of an

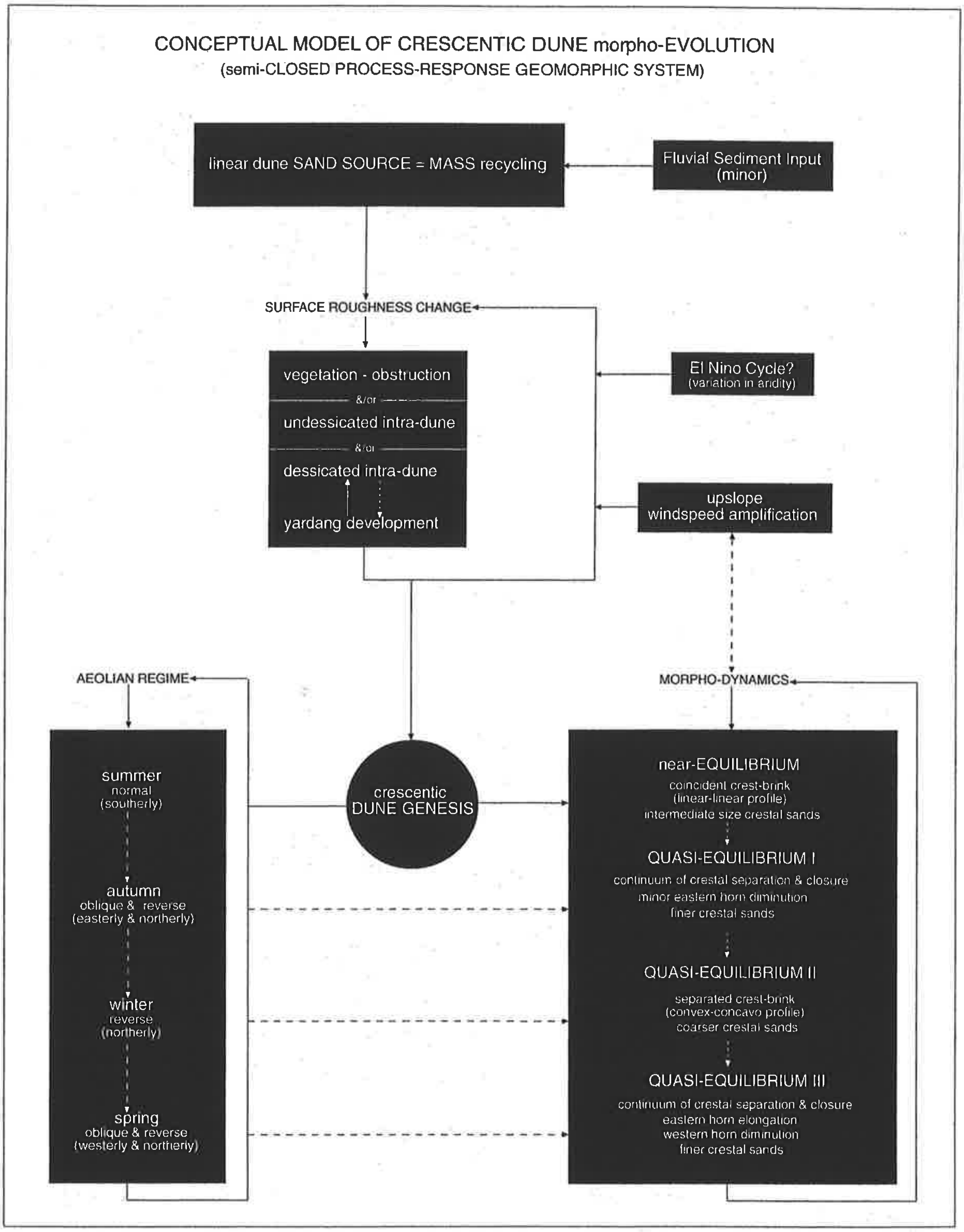

Figure 9.1 A conceptual model of the semi-closed geomorphic system at Gurra Gurra waterhole. Crescentic dune origin and development are shown to 'flow' from (i) an abundant but not overly copious sand supply, from which, (ii) changes in surface roughness of the clay-rich intra-dune allows sand patches to develop into barchan-like dunes of (iii) equilibrated form, that undergo 'whole-dune' sediment transport in the summer months of normal, unidirectional, high intensity air flow. The entire geomorphic system is seasonally cyclic and self-sustaining in that mass transfer is predominantly derived from the erosion of the linear dune with deposition into crescentic dunes.

inhomogenous, incongruous substrate distort dune morphology. Dimensional equilibrium is a transient feature at Gurra Gurra waterhole and not characteristic of this dunescape.

9.2 Surficial Sedimentology

The surficial sedimentology of the Gurra Gurra crescentic dunes is one of medium-to-fine quartz-rich sands with a unimodal, positively skewed and leptokurtic distribution. Analysis has shown that finer mean grain-sizes correlate with better sorting, a less positive skewness and lower kurtosis. This finding agrees with other studies (Lancaster, 1982b, 1987; McKee, 1983; Vincent, 1984) across the globe. The signatures of the sedimentological distributions across the dunes, and in different seasons, are subtle contrasts that are attributed to either the removal or accumulation of the fine tail of the distributions. Seasonal variation is gradual and transitional between each season, while greatest differences are found between the upper and lower micro-geomorphic positions of the dunes. Seasonal changes in dune shape simultaneously respond with transitional changes in surficial sedimentology. Both are responses to the seasonal change in aeolian regime. The processes that most contribute to the surficial sedimentology for the Gurra Gurra crescentic dunes involve:

- avalanching upon the lee and rear slipfaces;
- winnowing of intermediate-to-fine sand fractions;
- slope threshold effects;
- sifting and interstitial entrapment of fines.

Each sediment grain-size distribution is most probably an amalgam of these processes rather than the result of one specific action. It is very much a consequence of the available shear velocity of the upslope enhanced winds. For example, the contrast of surficial sedimentology between the upper and lower dune demonstrates the crest relative to the lower lee and stoss, to be coarsest in winter and late autumn. Granulometric differentiation by upslope threshold effects and the preferential winnowing of intermediate grade sands under the influence of reversed and upslope amplified winds upon the crest-brink, are responsible for such distributions. The transfer of winnowed materials between the upper and lower dune is a common feature of the sedimentology for these seasons, whilst the existence of slope threshold differentiated sand between the perimeter and crest of the dunes, results in greater distributional contrasts. However, the vehement normal-flow activity of the entire dune in late spring and summer counters the affects of upslope amplification of shear velocity, and assists in deriving a more turbulent mix of grain-sizes over a greater proportion of the dunes. In such circumstances, saltation of a broader grain-size range is the dominant mode of entrainment between dune toe and crest. Such a process lacks the subtlety of winnowing and interstitial sifting that are prevalent upon the upper dune when lower, upslope-amplified shear velocities are common, as in both autumn and winter.

In contrast, inter-seasonal variation of like geomorphic location is not significantly different for mean grain-size. This occurs, because it is the small volume of the positive tail of the distributions that most alters the sedimentological signature of the distributions for this exceptionally well refined sediment. The greater number of inter-seasonal contrasts for sorting and skewness concurs with this conclusion. Furthermore, without the comparison of slope-threshold differentiated sands of coarser (lower dune) and finer (upper dune) grades, a negligible contrast of mean grain-size exists between like micro-morphologic locations. Sedimentological contrast is not easy to identify when the sediment grain-size range is narrow. This explains why many more grain-size contrasts occurred between different morphologic positions compared to inter-seasonal contrasts of like-morphologic location.

It is also inferred that when the entire sand body has become *en masse* crescentic forms, or when the cycle of suspected El Niño aridity ceases, thereby extinguishing the supply of sand to the crescentic forms, the closed nature of this small dunescape assumes greater significance. Either the crescentic dunes will remain static with insignificant growth, and become stabilised vegetated features, or retrogress and degrade into incongruous sand sheets that cannot be maintained as dunes. Although it is uncertain which scenario will unfold, the assumption that crescentic dune activity will eventually cease appears to be justified. The proximal alluvial deposits of the modern Strzelecki Creek do not significantly contribute to the construction of the crescentic dunes. Although some alluvial sourced input of mass occurs by the oblique westerly winds of spring, the predominant aeolian transport direction is towards this source rather than away from it. This transport direction is inconsistent with continued dune development. Comparison of the sediment grain-size distributions of the anastomosing Strzelecki Creek alluvium with that of the crescentic dunes clearly distinguishes the bi-or-polymodal grain-size distributions of the fluvial sediments and the uni-modal signature of the aeolian sands. The presence of only minor quantities of coarser fluvial grain-sized sediment on the interdune and the periphery of the crescentic dunes, together with the absence of significant fine, fluvial translucent quartz sands within the dune bodies, attests to the limited mass transfer between the dune system and alluvial plain. This factor portends the eventual demise of the crescentic dunes, once the sediment source-dune is exhausted.

9.3 Aeolian Processes at Gurra Gurra: The Global Implications

It has been observed for many localities around the globe, that crescentic dunes in annual uni-directional aeolian regimes demonstrate both crest-brink separation and crest-brink coincidence (Bagnold, 1941; McKee, 1966; Hastenrath, 1967; Lancaster, 1985a, 1987; Havholm and Kocurek, 1988; Burkinshaw *et al.*, 1993). With inferences made from the modelling of the Gurra Gurra dunes, crest-brink separation in uni-directional currents normal to the crest, is a response to variation in the shear velocity *(u_*)* over the windward slope. Simply, this is a result of a changing pattern of erosion and deposition, as the point of maximum wind shear migrates under the influence of changing wind speeds. A decreasing shear velocity correlates with diminishing erosion. Subsequently, such diminution of u_* results in only crestal shear sculpting the upper dune. This causes a separation of the crest and brink, as seen by the deposition of the eroded crestal sands onto the upper lee, a 'rounding' and broadening of this region, and minor dune lowering. As crestal peakedness decreases with the subsequent leeward deposition of the crestal sands, the inclination of the slipface becomes less than the angle of repose for the upper lee. However, crest-brink separation diminishes the compression of flow lines and ultimately results in a morphological threshold where the cessation of crestal movement occurs, and a convex, slightly separated crestline is the dominant configuration of the dune. A quasi-equilibrated morphology is established in the new lighter wind regime. However, with the return of more intense shear velocities, which involve the entire windward slope in transport, crest-brink closure comes about with air flow constantly adjusting the entire sand mass through migration, and returns the dune to an equilibrated form. The form of the dunes become much more coherent or 'sharp', where the transfer of energy and mass is at an optimum balance. Such processes and subtle variations of crestal form were distinguishable only with the normal, uni-directional summer winds of Gurra Gurra. Overall, such adjustments typify the feedback response between slope angle, diurnal wind flow speeds, surface roughness and sand transport, while gross morphologic change and crestal reconfiguration was a consequence of multi-directional winds.

In accordance with the observations of Burkinshaw *et al.* (1993); Lindsay (1973); Smith (1970); Tsoar (1974) Sharp (1966) and Bagnold (1941), the work reported here, also demonstrates morphological diversity to be allied to the multi-directionality of the wind. Thus, the microcosm of crescentic dune genesis and evolution at Gurra Gurra waterhole reinforces the observations of greatest morpho-dynamic change occurring within a multi-directional, multi-intensity aeolian regime. Furthermore, the Gurra Gurra dunes emphasise that change is not haphazard, but part of a much larger environmental niche in which the dunes reside. It is also important to recognise that the absence of mathematical propriety of dune morphometry does not preclude the existence of systematic and genetic landform-scape development. Rather, the absence of simple mathematical relationships, as generally found in this work, infers a more complex operating system with a greater number of variables. Thus, whether the system be one of simplicity or complexity, a unifying model of process-response undoubtedly assists in giving a synopsis of the landscape under investigation (Fig. 9.1). When such an overview is available, it is then possible to produce comparative models of other dunescapes, which assist in identifying the common controls of form (equilibrium and quasi-equilibrium) and process, and the different interrelationships between site variables. For example, prevailing uni-directional winds produce a dune morphology in equilibrium, while the influence of a multi-directional wind regime delivers complex forms that only sustain dynamic equilibrium for a limited time. Although the dunes of Gurra Gurra are large enough to sustain crescentic form throughout each season, equilibrium of form is short-lived and only occurs for simple structures in the uni-directional and intense winds of summer. Otherwise a cycle of quasi-equilibrium is the norm (Fig. 9.1). Equilibrium of form is an abstract concept in landscape analysis, with even subtle crestal variation responding to diurnal changes in wind velocities, as earlier discussed. Morpho-dynamic equilibrium of crescentic dunes is represented by strong near perfect mathematical relationships that imply a simple environment of:

- a predominantly intense seasonal uni-directional air flow;
- a substrate of insignificant topographic variation;
- a relatively unvegetated and homogenous, cohesive pavement;
- a consistent but not overly copious sand supply.

If such parameters are maintained, then the process of morpho-dynamic equilibration concurs with the expression of ideal mathematical-morphometric signatures. However, such ideal conditions are infrequent if not rare, which results in dunes that are in a state of morpho-dynamic adjustment to the current environmental conditions - that is, quasi-equilibrium. Many stages of quasi-equilibrium can be seen within a dunescape, the greatest differences being attributed to the influence of multi-directional winds and the onset of gross morphological readjustments, with least differences being due to a varying wind speed in a uni-directional regime, and the more subtle changes that are afforded the crestal configuration of a dune (Fig. 9.1). Greater complexity and distance from morpho-dynamic equilibrium is further established when any of the other variables (topography, vegetation, pavement texture and sand supply) also depart from the ideal conditions discussed above.

Hence, it is apparent that the time of observation is crucial in defining the state of form. Dune systems, regardless of size, sand lithology and geographical location demonstrate a continuum of form and inheritance from preceding aeolian regimes. Only in environmental settings where confined unidirectional winds occur in all seasons, is a dune system maintained in morphodynamic equilibrium. It is here suggested that much of the disparity recorded for both morphology and sedimentology of like forms throughout the global deserts, is a result of observations being made within limited time intervals, that only represent a fraction of the total cycle in which a dunescape evolves. So-called controversies concerning the comparisons of form and sedimentology between global ergs, may simply be that observers fail to identify subtle seasonal contrasts in wind directions and intensities. This study of the Gurra Gurra crescentic dunefield demonstrated that regional sand drift potential was not effective at sand transport. However, site specific upslope wind speed enhancement was seen to be significant at changing the effective sand drift potential and, hence, crestal shape and surface sedimentology of a dune.

Throughout the literature, dune crests have been commonly reported as being finer, better sorted and less finely skewed than the lower slopes or plinths. This sedimentological signature is not unexpected, considering that sorting mechanisms operate more effectively and for longer periods on the crest than on the lower slopes. The grain-size distribution of crestal sands is simply a continual adjustment to variation in amplified upslope wind speeds. When the entire dune is active under intense winds, sorting is no more pronounced upslope than downslope. However, when the lower slopes of a dune are quiescent, crestal sorting can be prevalent, given sufficient regional wind energies. The absence of a crest displaying fine, well sorted, low skewed sands is evidence of either times of gross saltation by high energy winds over the entire dune, or, when reversed or oblique upslope relocation of sands occurs by selective grain-size winnowing in recently re-oriented winds.

Inheritance of sedimentology explains the absence of significant inter-seasonal contrasts in surficial sedimentological distributions for the Gurra Gurra dunes. Dunes under near morpho-dynamic equilibrium reflect the summer distributions. However, dunes that are in a state of quasi-equilibrium, portray sedimentologies that reflect the sediment budget (erosion, transport and deposition) under the influence of more minor reversing and oblique winds of varying intensity and duration. Due to the relatively low intensity of the autumn and winter winds at Gurra Gurra, it is the transport of fines to and from the upper dune onto the lower dune that constitutes the greatest sedimentological change. In contrast, the greater energies of the summer and spring winds signify sand entrainment from the lower dune onto and over the upper dune, and the times of greatest dune mobility and migration. Hence, once the crescentic form of the dunes is established under the most vehement wind conditions, the essential dune sedimentology is established. Any changes that follow will usually be less significant when compared to the formative conditions and will be revealed as subtle changes to the removals and accumulations of the tails of the sedimentary distributions and the morphological reconfiguration of the upper dune. However, secondary wind conditions that approach the intensity of those that form the dunes, will result in far greater morphological reconfiguration and sedimentological flux. For example, a reversing secondary wind of strength and duration equal to the primary aeolian regime may be of such intensity and duration that it causes a total reversal of form, and a sedimentology that reflects sand movement over the entire dune. It is evident that dune sedimentology and morphology reflects periodicity, and hence predictability with the seasonal aeolian cycle.

9.4 Research Methods: A Retrospective View

The measurement of landform dimensions is fundamental to geomorphology. The methods of mathematical analysis used here, were characteristic methods of quantitative geomorphology. The greatest consideration that was required in the morphometric examination of the dune data sets, pertained to the decision to use either parametric or non-parametric statistical tests.

In contrast to the morphometric methods, the resolution of dune sedimentology required considerable attention and fore-thought about which techniques to use. As reported in Chapter 3, both field sampling and laboratory preparation of dune sands have historically used a multitude of sampling techniques, sampling weights, sampling depths, and laboratory and statistical methods. Although sedimentological methods have rapidly evolved in the past few decades, no single technique nor style of analysis has been found to offer acceptable levels of accuracy and reliability in all circumstances in which desert sands are analysed (refer to Chapter 3 for a full discussion of the methods deemed most appropriate to this study). Nonetheless, the results here given, are extremely precise and are believed to be both representative of the aeolian sedimentological distributions, and informative about the processes acting on the Gurra Gurra dunes. The distribution-free Kruskal-Wallis and Scheffé-type projection methods are optimal at deducing differences of sedimentology between comparative micro-morphologic sites.

Nonetheless, probable sources of experimental error in this type of study are seen to be a consequence of inadequate dune sample numbers or sand sample numbers.

Errors can be attributed to:

- the small number of dunes sampled;
- the intrinsic variation in dune size, which gives variation in both slope transport distances and upslope shear velocities;
- the 'smearing or blending effect' of moments, when only upper and lower micro-morphologic locations were compared with a succession of upslope sediment samples, or when too few sand samples were taken within numerous micro-morphologic partitions.

With only small differences characterising the aeolian sediments of the crescentic dunes of Gurra Gurra, the examination of (possibly) too few micro-morphologic elements and/or (possibly) too few sand samples could (possibly) have precluded identification of the more subtle changes.

Such potential problems could be minimised by a sampling procedure based upon:

- single sample points at the bases of the stoss and lee, and crest for a very large sample of dunes ($n \geq 30$) of near identical dimensions (the influence of slight differences in dimensional variation and any slope enhanced grain-size differentiation will become negligible as the dune sample size increases);
- subdivision of the dune into (5) or (6) micro-morphologic partitions where the number of dunes can be smaller than that mentioned above, but the number of sand samples within each partition must be large.

The number of sand samples will depend upon dune dimensions. Numbers of ($3<n<5$) for each partition along a longitudinal profile, should be a reasonable estimate for dunes the size of those at Gurra Gurra. Such sample numbers would also offer themselves to rigorous statistical examination. Equally, the elimination of any 'smearing effect' for samples near partition boundaries would be less evident.

Consequentially, sedimentological comparisons over equal transport distances, but not like-micro-morphologic location for dunes of different dimensions, can also be made, and aid in determining whether a threshold of upslope size differentiation of surface sediment occurs with dunes of a specific dimension. Although identification of this would be difficult to ascertain, due to the influence of upslope wind speed amplification and the increase in u_* and sand mobility that occurs. This is a further reason to choose dunes of near identical dimensions when the objective is to deduce statistically likely process-controls from the sedimentological distributions. Thus, to eliminate any possible upslope variation of process and grain-size that may intrinsically occur between larger, taller dunes relative to smaller, lower structures, only dunes of similar size should be compared.

Each of the above alternatives, will demonstrate sedimentological change over a crescentic dune. However, the second alternative is a better indicator of transitional change. When combined with a large sample of dune numbers and the procedures outlined in Chapter 3, this method will derive the most accurate record of localised sedimentological change through morphodynamic processes. It is pertinent to note that previous studies around the globe have, in most cases, only examined single dunes or only single seasons (even when multiple dunes samples are employed), and cannot be regarded as definitive accounts of dune evolution. Rather, these analyses are indiscriminate suggestions of change that demonstrate the need for detail, and emphasise that both spatial and temporal change are important in defining an understanding of dune form, sedimentology and process.

9.5 Implications and Future Research Directions

In conclusion, this dissertation has examined the spatial and temporal geomorphology and surficial sedimentology of aeolian bedforms within a modest, semi-closed geomorphic system. To date, this appears to be the first comprehensive study of its type on Australian desert dunes. Likewise, the use of the Kruskal-Wallis test and a Scheffé-type projection are analytical techniques that have not been previously employed to compare desert dune sedimentology. In the work reported here, such techniques were highly successful at discerning contrasts between the sedimentological signatures of dune-sand samples and would assist in future studies of dune sedimentology.

The study of the Gurra Gurra dunes has also enlightened the understanding of the global contrasts found with dune morphology and sedimentology. This research has demonstrated and emphasised the cyclical response of aeolian landforms to often subtle changes in seasonal conditions and the necessity of long-term, periodic observations. Without these considerations, the explanation of dune configuration and sedimentology for the many and varied global arid zones will always remain ambiguous.

A primary purpose of this study was to attempt to identify the processes responsible for the morphological and sedimentological variation of the Gurra Gurra dunes. This thesis has explored the mechanism of process control of aeolian landforms, and has succeeded at integrating the variables of form and sedimentology with process control. Generally, process models have not been a feature of prior studies. Similarly, links between erosional landforms, depositional landforms and processes have been made here. Without the holistic examination of the local landscape at both the micro- and meso-scales, as well as, the interrelationships between erosion and deposition, it would not have been possible to deduce the manner in which the Gurra Gurra dunes originated and developed.

Importantly, the investigation of yardang evolution at Gurra Gurra has questioned the likelihood of clay-enriched desiccated substrates on Mars. The existence of such features for past fluvial and lacustrine terrains, may provide a mechanism from which sand-sized particles have originated from predominantly basaltic lithologies and were later sourced into Martian dunes. The processes of dune-building and the genesis of sand-sized particles on Mars, have long been an issue of controversy. The link between dunes, yardangs and desiccated surfaces requires much further consideration in a Martian context, and may offer some insight into the genesis of localised Martian aeolian landforms.

Finally, the time-honoured geomorphological tradition of differentiating landform equilibrium from landform quasi-equilibrium has been used to explain morphological change. Such an approach has allowed integration of the factors of dune form, geographical setting and the mathematical relationships produced by these factors. In doing this, the identification of equilibrated simple dunes for the Arequipa plain of Peru and the predominantly quasi-equilibrated compound dunes of Gurra Gurra waterhole have shown that the many occurrences of equilibrated crescentic dunes elsewhere, correlate with environmental conditions that reflect dominant unidirectional winds, a congruous substrate without gross topographical variation and a regular but only subsistence sand supply. Such a proposal has definitive value in describing the aeolian regime of Mars. The absence of linear dunes, except for localised examples of seif-linear segments, further characterises the simple climatological and meteorological environments that have occurred across a water deficient planetary surface.

Future research directions in aeolian studies must reconsider fundamental principles such as the sedimentological sampling technique, sample size, and laboratory preparation and analysis methods. Too many so-called 'tried and true' methods are employed without consideration of the compatibility of the research objective and technique. The application of the log-hyperbolic distribution needs further enquiry into its foreseen advantages over log-normal methods, as do the methods of granulometric analysis and the derivation of descriptive statistics for both uni-modal and especially poly-modal distributions. To date, there is no adequate method that could be considered incontrovertible at deriving descriptive statistics and the environmental meaning of bi- or tri-modal sedimentological distributions.

Similarly, it is no longer acceptable to simply identify morphological differences of landforms without, at the same time, quantitatively attempting to deduce the processes responsible for shaping the landscape. If this is not done, the plethora of measurement, statistical and observational data has little significance and can justifiably be considered as mere 'stamp collecting'. Future modelling must incorporate measurements of dune enhanced wind speeds and directions, and the measurement of substrate cohesiveness and its role in the rate of dune re-equilibration within a particular wind direction and intensity. Engineering practices of determining soil moisture content and soil compressibility may assist in this matter. Coupled with statistical studies of grain morphology, and the relationships between grain-size and roundness, and transport potential, these will better define the overall granulometric characteristics and formative mechanisms of dunes.

This dissertation has contributed to a basic understanding of the manner in which the wind sculpts the terrestrial arid landscape. However, the future use of extra-terrestrial analogues, such as the desert landscapes of Mars, are indispensable in producing further insights into wind as a geological process. The aridity of Mars displays analogues of landforms of immense areal coverage and dimensions that often extend far beyond those of terrestrial terrains. In many respects the landforms of Mars are 'frozen' in time, although the chronology of much of the aeolian landscape is currently ambiguous. Pristine planetary landscapes that no longer exist on Earth due to its dynamic resurfacing, are available for the testing and comparison of evolutionary models. If nothing else, extra-terrestrial landscapes offer a larger geomorphic inventory from which models can be made and compared under different atmospheric, compositional and gravitational variables. Planetary landscapes are more than an enlarged data base. Breed's (1977) finding of geometrical similarity between the crescentic dunes of Earth and Mars, offers both a scientific eloquence for the universal laws of nature beyond the comfortable terra-centric reference frame in which so many geologists-geomorphologists reside, as well as an inordinate esoteric beauty which only landscape analysis (geomorphology) can give. Thus, rather than considering extra-terrestrial environments as exotic, unfamiliar annexes to terrestrial themes, far greater use should be made of what these distant landscapes have to offer.

This dissertation has revealed through the integration of qualitative and quantitative process-geomorphology that the Strzelecki-Gurra Gurra dunescape portrays a microcosm of aeolian features, and processes, that are not dissimilar to many in the immense areal deserts of Earth and Mars. It is hoped that the observations and discussions here given, benefit the understanding of spatial and temporal change and the genesis and evolution of aeolian landforms.

Dune Number	Length (m)	Width (m)	Wavelength (m)	Type
1	35.0	21.7	------	BD
2	32.5	26.7	50.0	BN
3	35.0	31.7	36.7	BN
4	15.0	18.3	32.5	B
5	20.8	13.3	36.6	BN
6	32.5	20.0	32.5	BN
7	22.5	19.2	------	BN
8	22.5	13.3	31.7	SE
9	35.0	22.5	------	BE
10	22.5	21.7	25.8	BN
11	29.2	23.3	43.3	BN
12	20.0	15.8	42.5	BNS
13	25.8	26.7	45.0	BN
14	13.3	10.0	31.7	BN
15	30.0	15.0	------	SE
16	26.7	20.0	46.7	BE
17	26.7	29.2	42.5	BNS
18	24.2	20.8	50.8	BN
19	33.3	20.0	23.3	BN
20	27.5	15.0	------	BN
21	29.2	33.3	46.7	BN
22	26.7	17.5	48.3	BE

Appendix 1 Planimetric Dimensions of the Strzelecki Creek Crescentic Dunes (1984)

KEY

B individual barchan

BD domical morphology with barchan-like slipface

BE individual barchan with elongation of horn

BN barchanoid

SE seif-like elongation with a barchan-style slipface

BNS barchanoid with seif-like elongation

Dune Number	Length (m)	Width (m)	Wavelength (m)	Type
23	24.2	27.5	36.7	BE
24	26.7	23.3	28.3	BNS
25	18.3	22.5	34.2	BNS
26	29.2	25.0	------	BN
27	22.5	19.2	------	SE
28	33.3	25.0	------	BE
29	27.5	17.5	------	BN
30	23.3	26.7	49.2	BN
31	21.7	23.3	29.2	BN
32	33.3	28.3	35.8	BN
33	23.3	20.0	35.0	BN
34	------	10.8	45.8	BN
35	32.5	20.0	47.5	BN
36	25.0	23.3	------	BNS
37	36.7	16.7	------	BNS
38	37.5	25.0	61.7	BN
39	37.5	18.3	------	BN
40	37.5	28.3	46.7	BNS
41	35.0	45.0	-------	BNS
42	28.3	22.5	-------	BN

Appendix 1 Planimetric Dimensions of the Strzelecki Creek Crescentic Dunes (1984)

KEY

B individual barchan
BD domical morphology with barchan-like slipface
BE individual barchan with elongation of horn
BN barchanoid
SE seif-like elongation with a barchan-style slipface
BNS barchanoid with seif-like elongation

Dune Number	Width (m)	Slipface Length (m)	Slipface Inclination (θ°)
1	18.70	7.80	32
2	21.20	3.40	32
3	14.00	4.90	32
4	24.70	5.10	31
5	19.10	4.85	32
6	18.50	3.90	32
7	30.40	7.70	31
8	18.90	4.90	31
9	28.10	9.40	31
10	17.80	5.80	32
11	46.30	11.20	32
12	15.80	3.10	31
13	18.30	12.20	32
14	23.00	5.10	32
15	26.60	7.30	31
16	44.90	7.60	32
17	51.10	9.70	32
18	16.40	6.00	30
19	23.60	8.70	32
20	11.80	9.40	30
21	25.50	8.20	31
22	24.60	4.90	33

Appendix 2 Morphometry of the Gurra Gurra Crescentic Dunes (early Autumn 1993). Dune numbers <u>do not</u> correspond with those in Appendix 1.

Dune Number	Width (m)	Slipface Length (m)	Slipface Inclination (θ°)
23	15.80	6.20	34
24	15.90	4.70	30
25	42.80	4.00	31
26	13.00	8.80	32
27	18.40	5.90	32
28	29.30	5.80	30
29	16.10	6.70	32
30	18.30	6.40	31
31	31.00	5.00	30

Appendix 2 cont. Morphometry of the Gurra Gurra Crescentic Dunes (early Autumn 1993). Dune numbers do not correspond with those in Appendix 1.

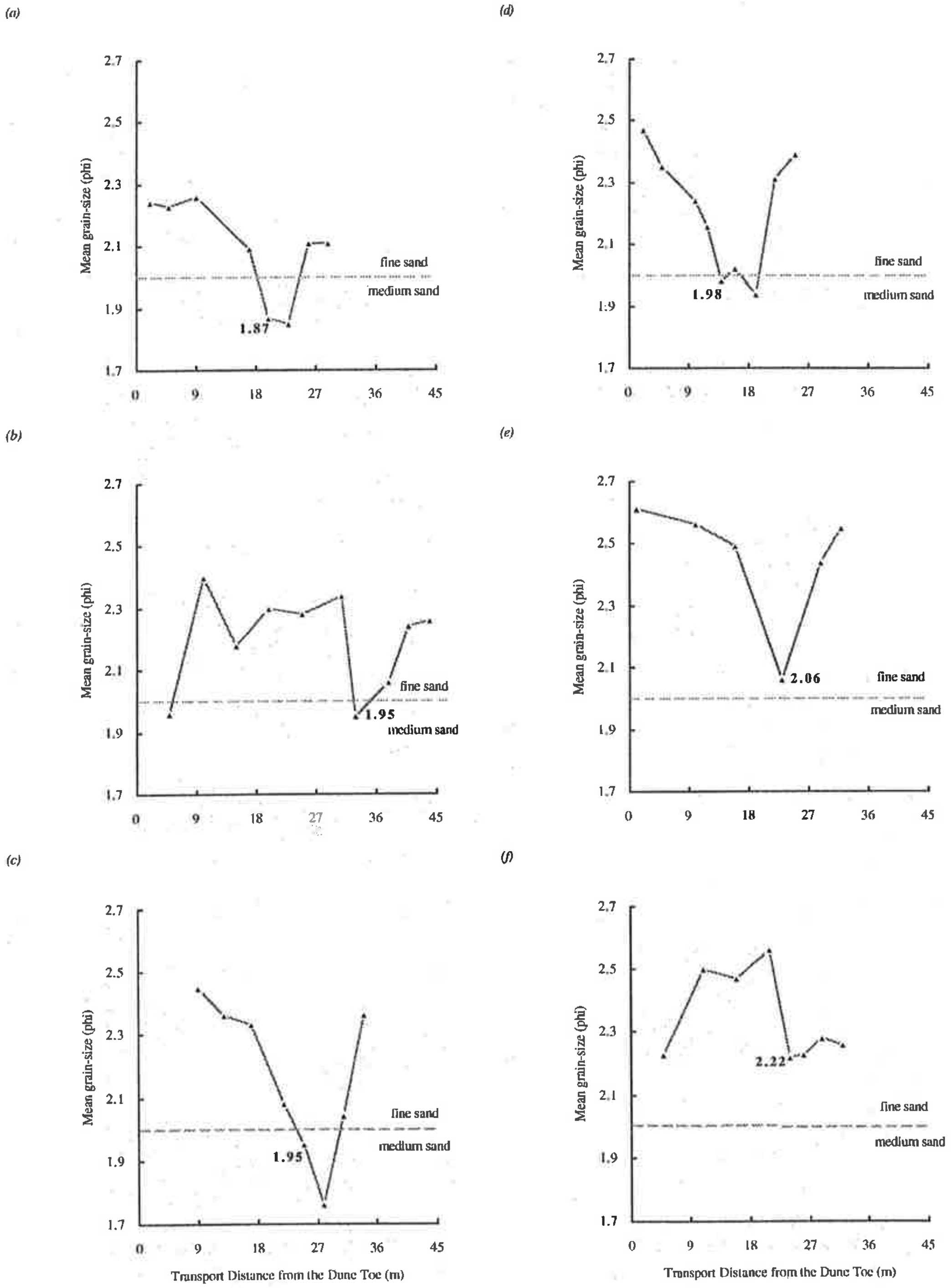

Figure 6.4 Longitudinal profiles of mean grain-size from the dune toe (lower stoss) towards the dune lee, for dunes (1 - 6) in winter 1991. (Bold numerals represent the mean grain-size of the crest).

DESCRIPTIVE STATISTICS FOR THE FIRST MOMENT (MEAN GRAIN-SIZE - ϕ units)

1. LEE
- $n = 18$
- median: 2.24
- maximum: 2.55
- minimum: 1.76
- range: 0.79

2. CREST
- $n = 6$
- median: 1.97
- maximum: 2.22
- minimum: 1.87
- range: 0.35

3. STOSS
- $n = 25$
- median: 2.33
- maximum: 2.61
- minimum: 1.96
- range: 0.65

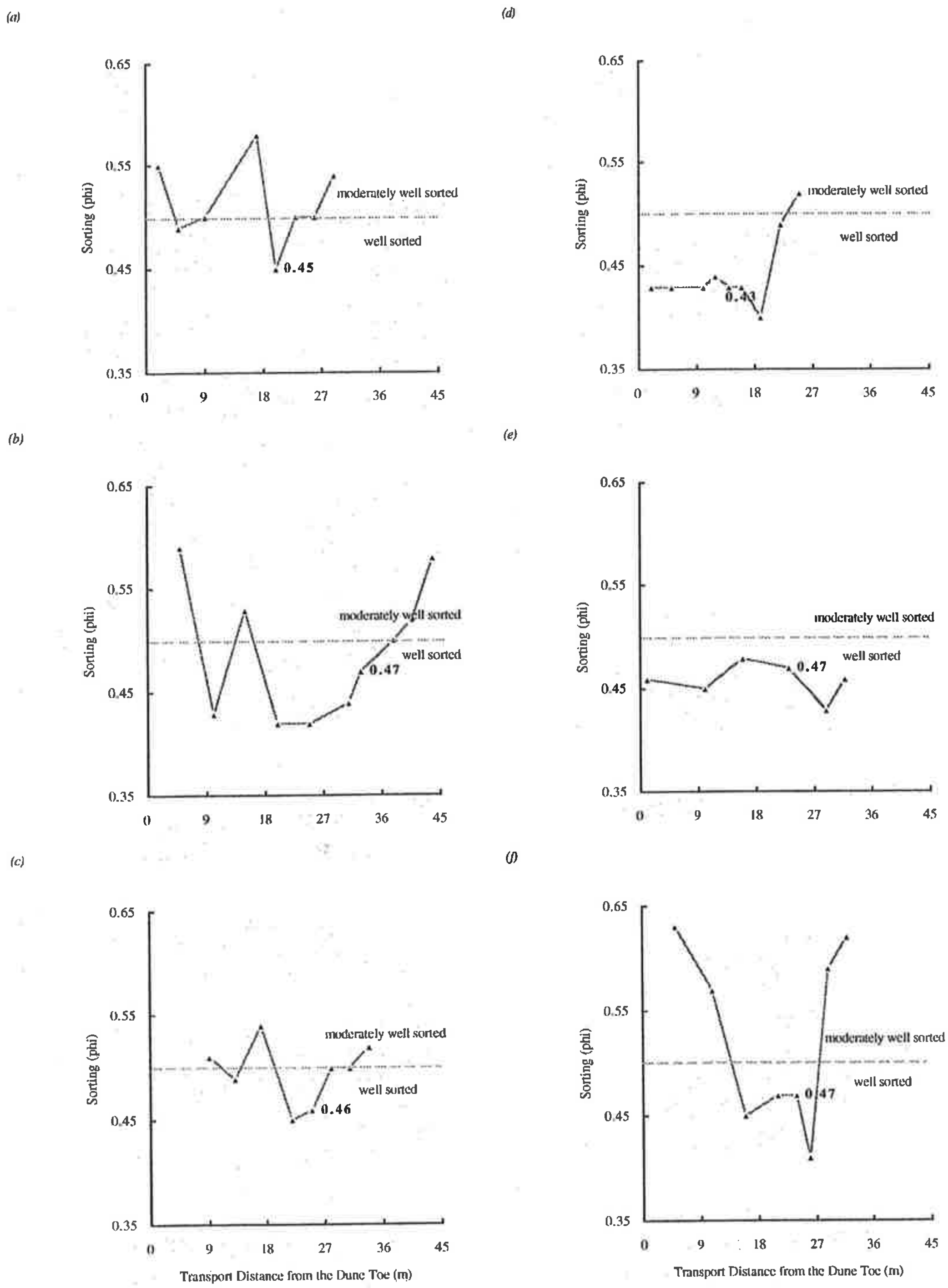

Figure 6.5. Longitudinal profiles of sorting from the dune toe (lower stoss) towards the dune lee, for dunes (1 - 6) in winter 1991. (Bold numerals represent the standard deviation (sorting) of the crest).

DESCRIPTIVE STATISTICS FOR THE SECOND MOMENT (SORTING)

1. LEE
- $n = 18$
- median: 0.50
- maximum: 0.62
- minimum: 0.40
- range: 0.22

2. CREST
- $n = 6$
- median: 0.47
- maximum: 0.47
- minimum: 0.43
- range: 0.04

3. STOSS
- $n = 25$
- median: 0.47
- maximum: 0.63
- minimum: 0.42
- range: 0.21

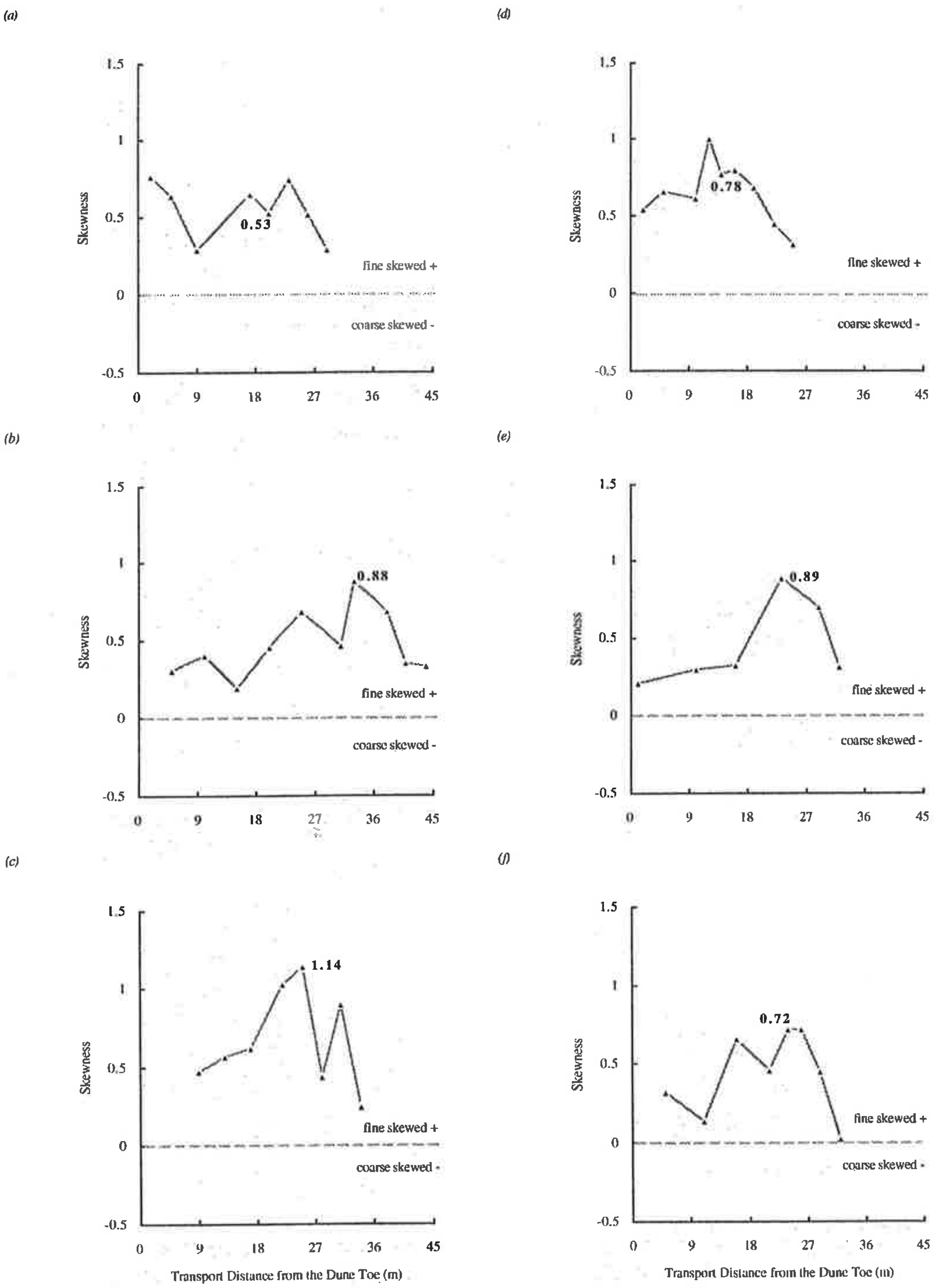

Figure 6.6. Longitudinal profiles of skewness from the dune toe (lower stoss) towards the dune lee, for dunes (1 - 6) in winter 1991. (Bold numerals represent the third moment (skewness) of the crest).

DESCRIPTIVE STATISTICS FOR THE THIRD MOMENT (SKEWNESS)

1. LEE
- $n = 18$
- median: 0.45
- maximum: 0.90
- minimum: 0.03
- range: 0.87

2. CREST
- $n = 6$
- median: 0.83
- maximum: 1.14
- minimum: 0.53
- range: 0.61

3. STOSS
- $n = 25$
- median: 0.47
- maximum: 1.03
- minimum: 0.14
- range: 0.89

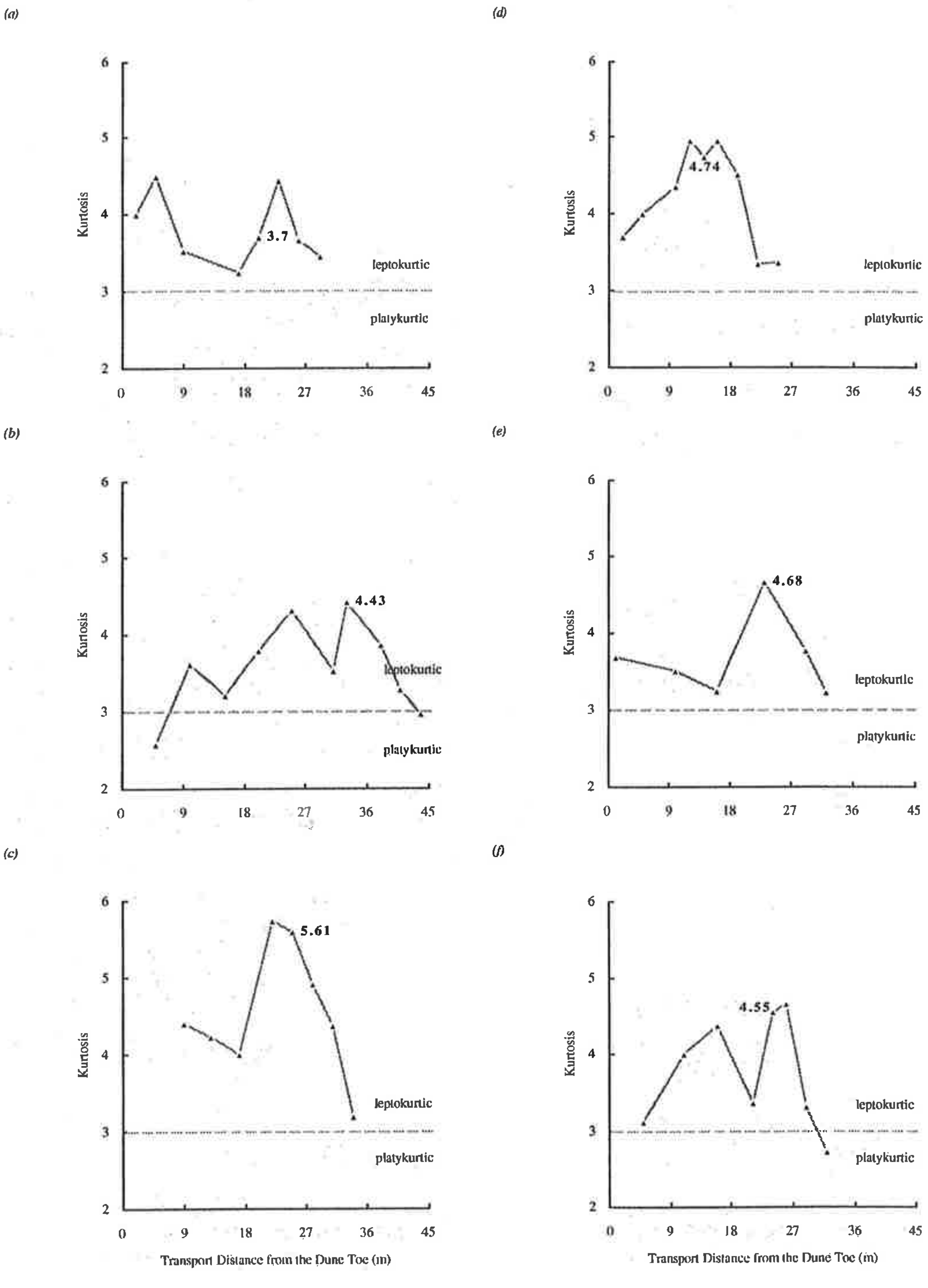

Figure 6.7. Longitudinal profiles of kurtosis for dunes (1 - 6) from the dune toe (lower stoss) towards the dune lee, in winter 1991. (Bold numerals represent the fourth moment (kurtosis) of the crest).

DESCRIPTIVE STATISTICS FOR THE FOURTH MOMENT (KURTOSIS)

1. LEE
- $n = 18$
- median: 3.57
- maximum: 4.96
- minimum: 2.72
- range: 2.24

2. CREST
- $n = 6$
- median: 4.62
- maximum: 5.61
- minimum: 3.70
- range: 1.91

3. STOSS
- $n = 25$
- median: 3.79
- maximum: 5.75
- minimum: 2.59
- range: 3.16

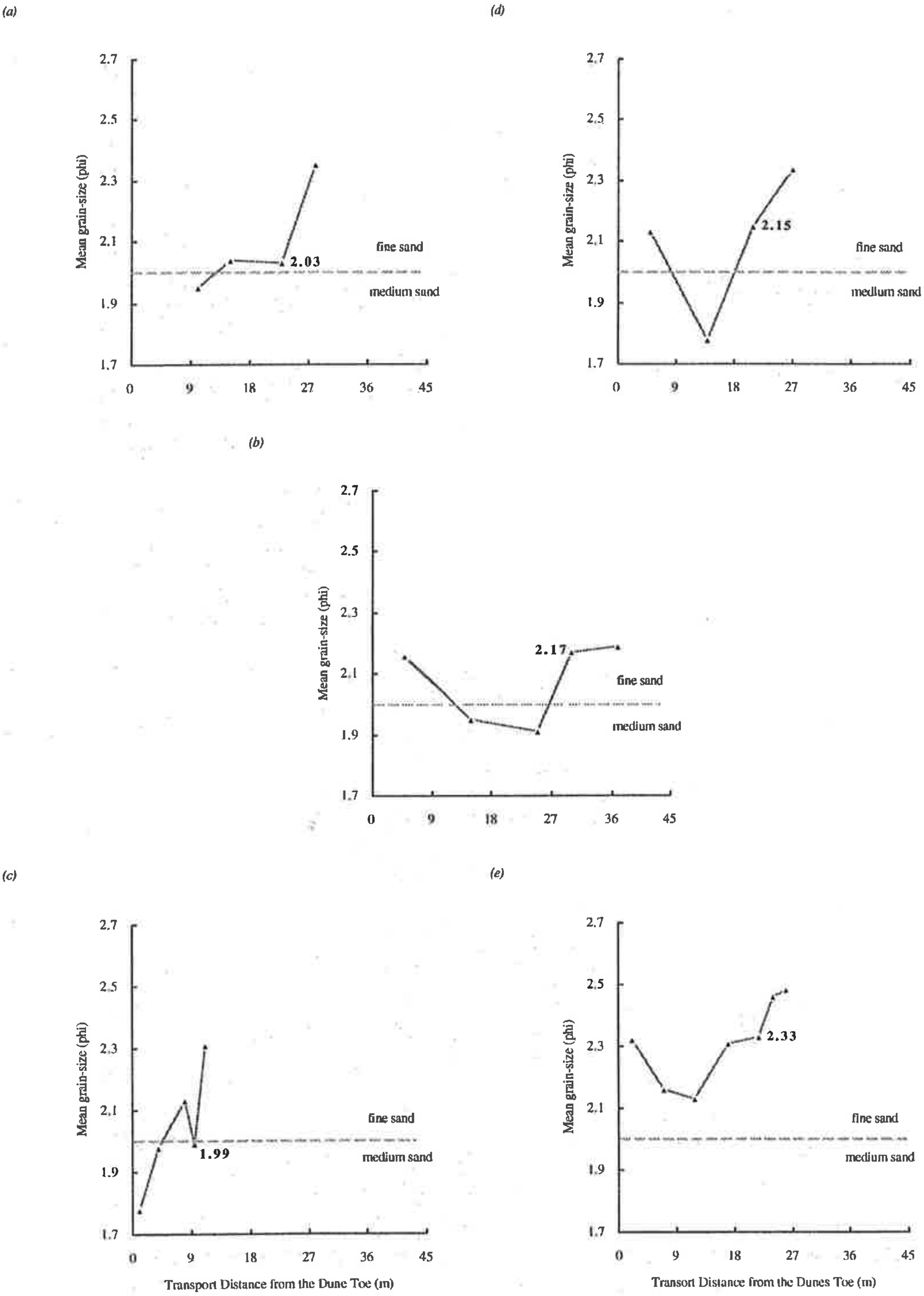

Figure 6.8. Longitudinal profiles of mean grain-size from the dune toe (lower stoss) towards the dune lee, for dunes (1, 2, 3a, 3b, 6) in summer 1992. (Bold numerals represent the mean grain-size of the crest).

DESCRIPTIVE STATISTICS FOR THE FIRST MOMENT (MEAN GRAIN-SIZE - ϕ units)

1. LEE
- $n = 6$
- median: 2.35
- maximum: 2.48
- minimum: 2.19
- range: 0.29

2. CREST
- $n = 5$
- median: 2.15
- maximum: 2.33
- minimum: 1.99
- range: 0.34

3. STOSS
- $n = 14$
- median: 2.09
- maximum: 2.32
- minimum: 1.78
- range: 0.54

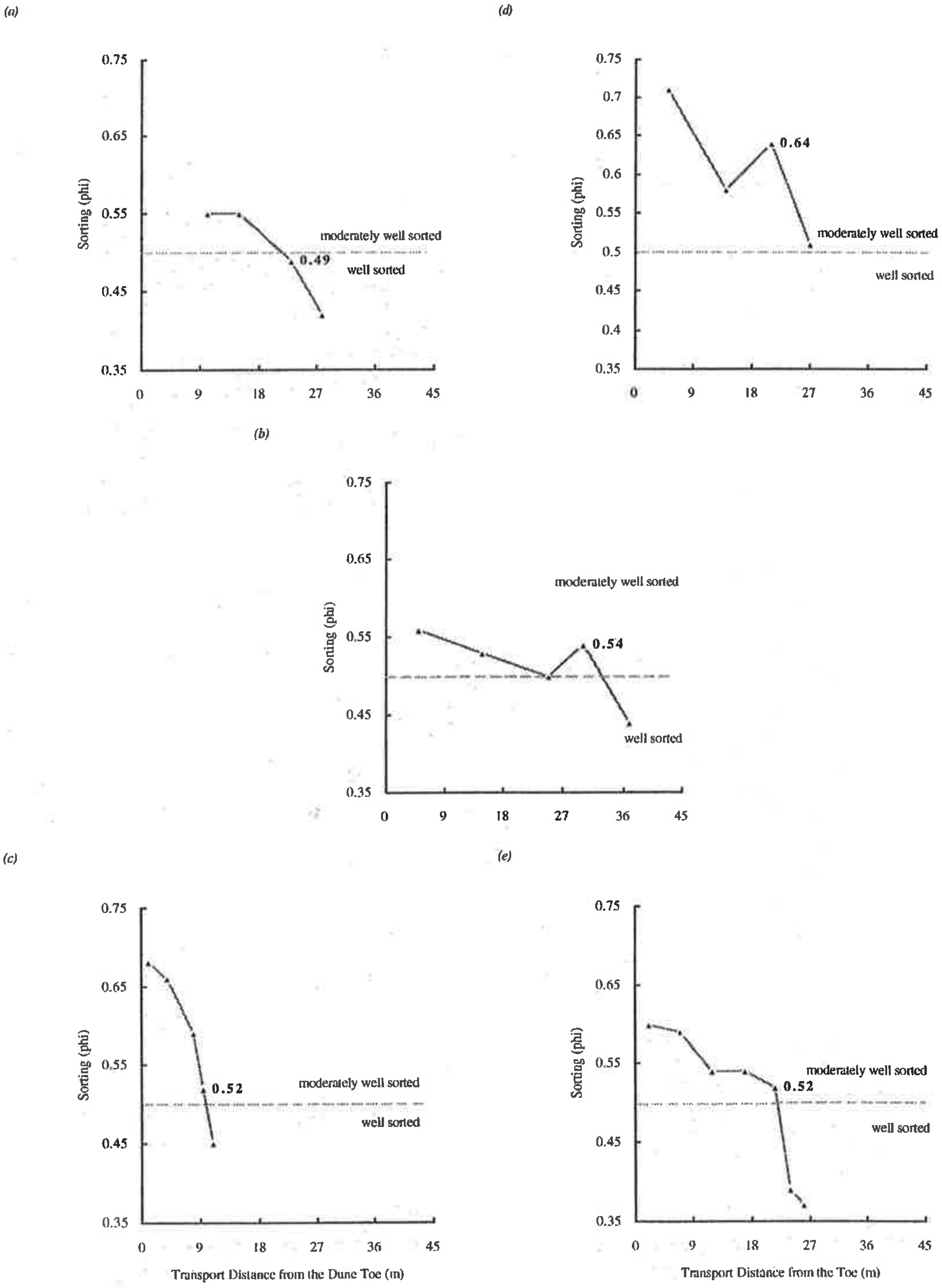

Figure 6.9. Longitudinal profiles of sorting from the dune toe (lower stoss) towards the dune lee, for dunes (1, 2, 3a, 3b, 6) in summer 1992.
(Bold numerals represent the standard deviation (sorting) of the crest).

DESCRIPTIVE STATISTICS FOR THE SECOND MOMENT (SORTING)

1. LEE
- *n* = 6
- median: 0.43
- maximum: 0.51
- minimum: 0.37
- range: 0.14

2. CREST
- *n* = 5
- median: 0.52
- maximum: 0.64
- minimum: 0.49
- range: 0.15

3. STOSS
- *n* = 14
- median: 0.57
- maximum: 0.71
- minimum: 0.50
- range: 0.21

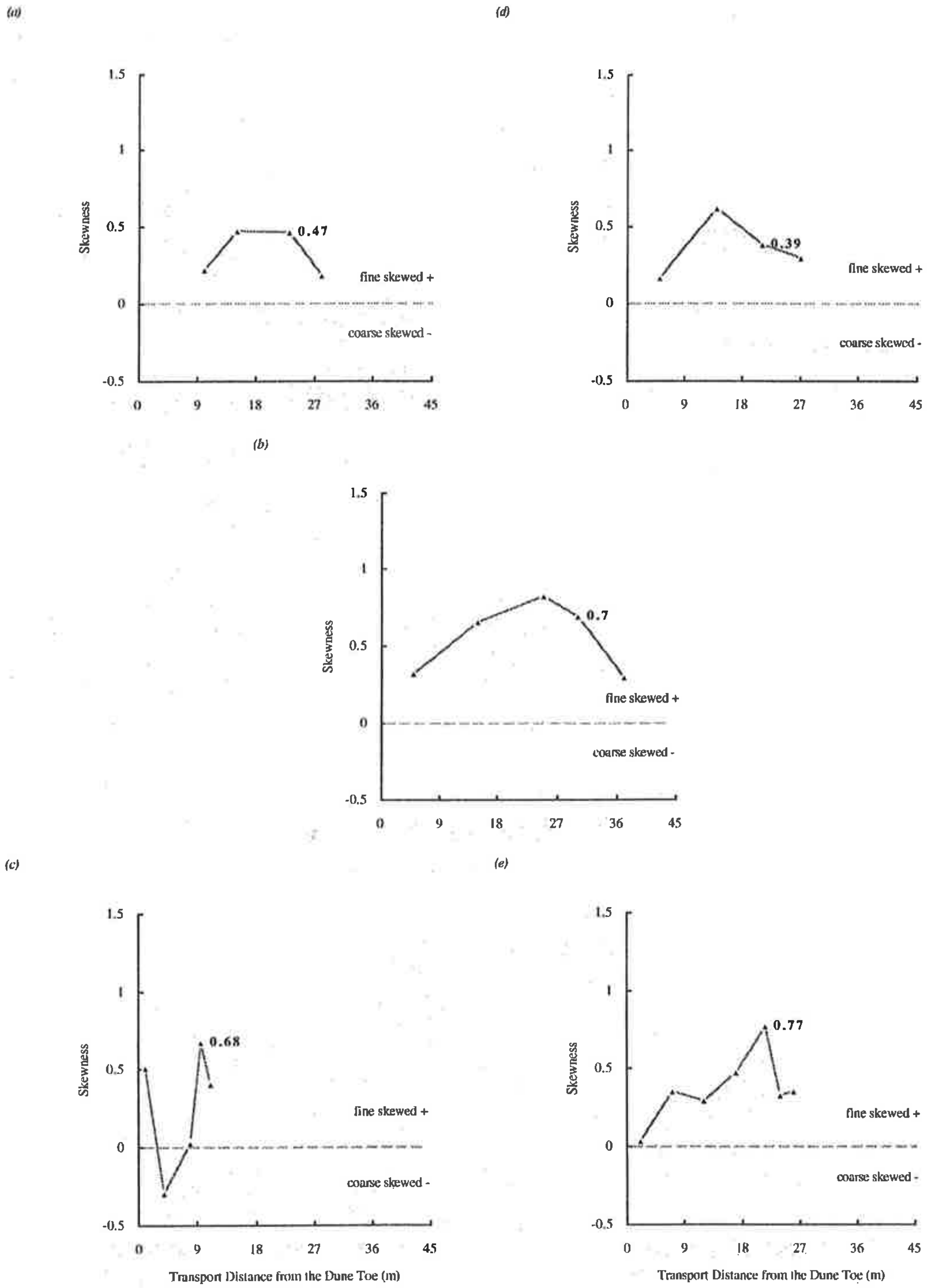

Figure 6.10 Longitudinal profiles of skewness for the sediment samples of dunes (1, 2, 3a, 3b, 6), from the dune toe (lower stoss) towards the dune lee in su
(Bold numerals represent the third moment (skewness) of the crest).

DESCRIPTIVE STATISTICS FOR THE THIRD MOMENT (SKEWNESS)

1. LEE
- $n = 6$
- median: 0.32
- maximum: 0.41
- minimum: 0.19
- range: 0.22

2. CREST
- $n = 5$
- median: 0.68
- maximum: 0.77
- minimum: 0.39
- range: 0.38

3. STOSS
- $n = 14$
- median: 0.35
- maximum: 0.83
- minimum: -0.29
- range: 1.12

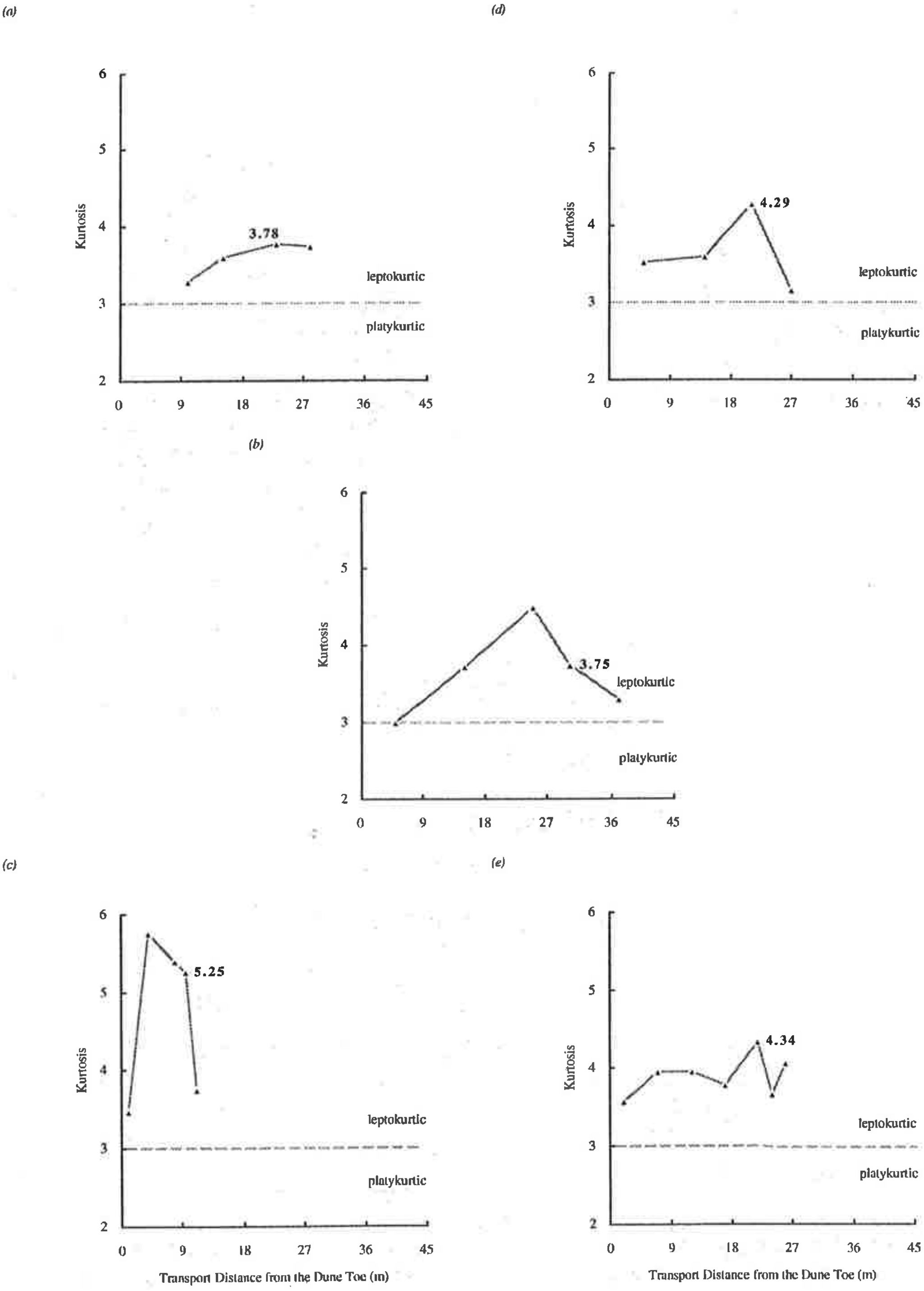

Figure 6.11 Longitudinal profiles of the kurtosis for the sediment samples of dunes (1, 2, 3a, 3b, 6), from the dune toe (lower stoss) towards the dune lee in s
(Bold numerals represent the fourth moment (kurtosis) of the crest).

DESCRIPTIVE STATISTICS FOR THE FOURTH MOMENT (KURTOSIS)

1. LEE
- $n = 6$
- median: 3.70
- maximum: 4.06
- minimum: 3.16
- range: 0.90

2. CREST
- $n = 5$
- median: 4.29
- maximum: 5.25
- minimum: 3.75
- range: 1.50

3. STOSS
- $n = 14$
- median: 3.67
- maximum: 5.76
- minimum: 3.01
- range: 2.75

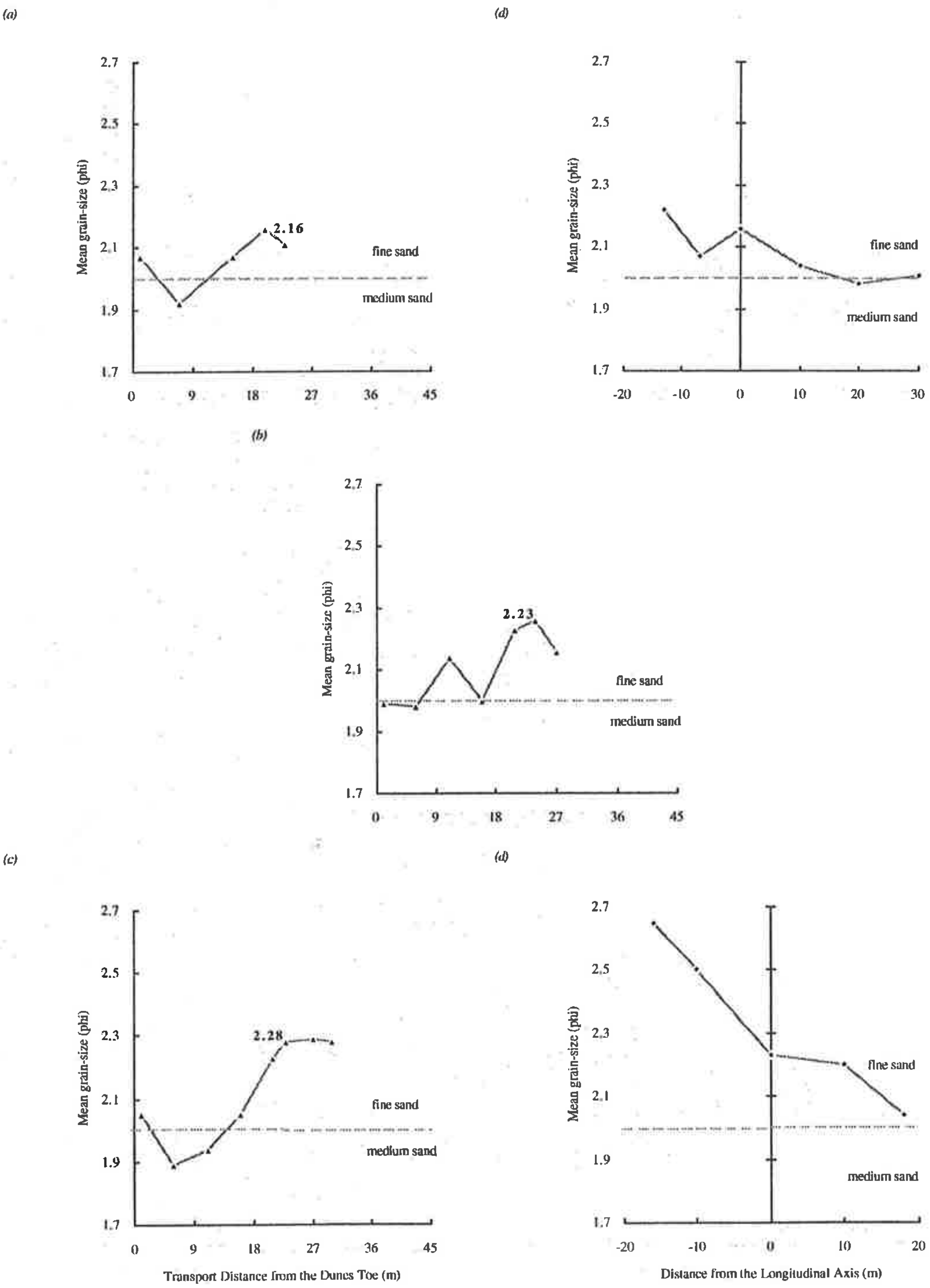

Figure 6.12 Longitudinal profiles *(a - c)* of mean grain-size from the dune toe (lower stoss) towards the dune lee, for dunes (3, 7, 8) and transverse profiles of dunes (3 & 7) in autumn 1993. (Bold numerals represent the mean grain-size of the crest while negative *x* values represent the western horn, and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE FIRST MOMENT (MEAN GRAIN-SIZE)

1. LEE

- *n* = 5
- median: 2.26
- maximum: 2.29
- minimum: 2.11
- range: 0.18

2. CREST

- *n* = 3
- median: 2.23
- maximum: 2.28
- minimum: 2.16
- range: 0.12

3. STOSS

- *n* = 12
- median: 2.03
- maximum: 2.23
- minimum: 1.89
- range: 0.34

4. WESTERN HORN

- *n* = 4
- median: 2.36
- maximum: 2.65
- minimum: 2.07
- range: 0.58

5. EASTERN HORN

- *n* = 5
- median: 2.04
- maximum: 2.20
- minimum: 1.98
- range: 0.22

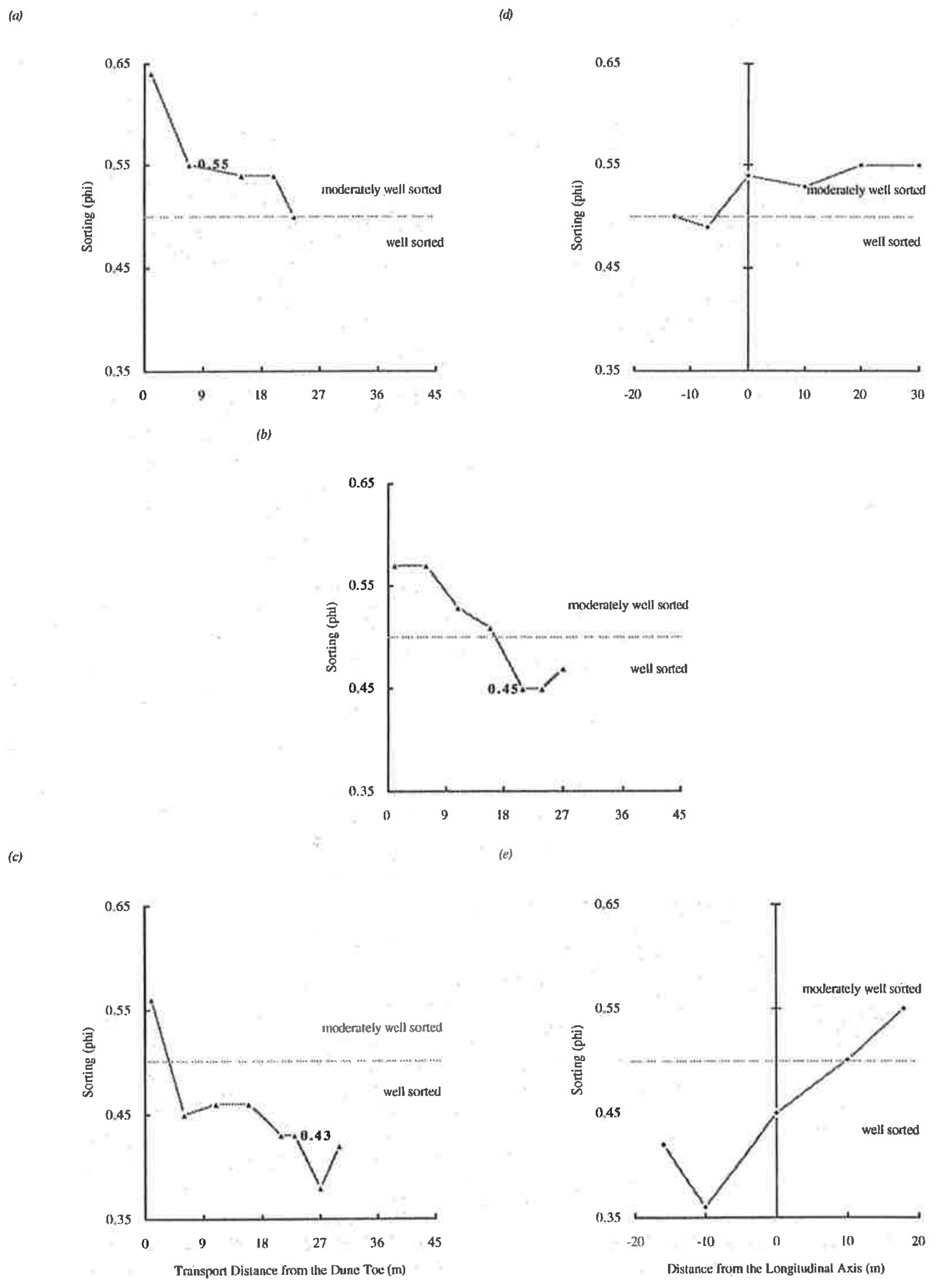

Figure 6.13 Longitudinal profiles *(a-c)* of sorting from the dune toe (lower stoss) towards dune lee for dunes (3, 7, 8), and transverse profiles of dunes (3 & 7) in autumn 1993. (Bold numerals represent the standard deviation (sorting) of the crest, while negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE SECOND MOMENT (SORTING)

1. LEE
- *n* = 5
- median: 0.45
- maximum: 0.50
- minimum: 0.38
- range: 0.12

2. CREST
- *n* = 3
- median: 0.45
- maximum: 0.54
- minimum: 0.43
- range: 0.11

3. STOSS
- *n* = 12
- median: 0.54
- maximum: 0.64
- minimum: 0.43
- range: 0.21

4. WESTERN HORN
- *n* = 4
- median: 0.46
- maximum: 0.50
- minimum: 0.36
- range: 0.14

5. EASTERN HORN
- *n* = 5
- median: 0.55
- maximum: 0.55
- minimum: 0.50
- range: 0.05

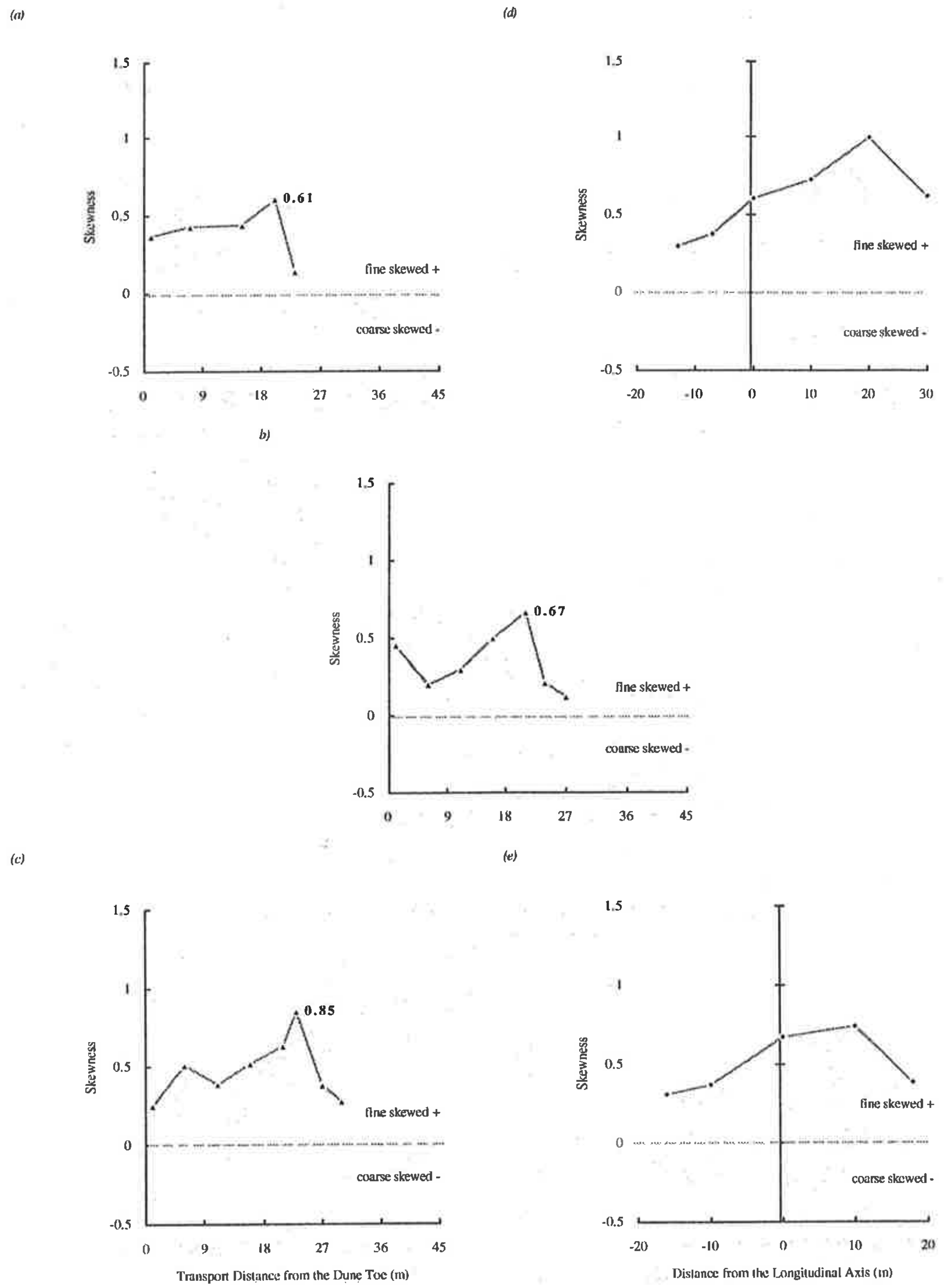

Figure 6.14 Longitudinal profiles *(a - c)* of skewness from the dune toe (lower stoss) towards the dune lee, for dunes (3, 7, 8) and transverse profiles of dunes (3 & 7) in autumn 1993. (Bold numerals represent the skewness of the crest while negative *x* values represent the western horn, and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE THIRD MOMENT (SKEWNESS)

1. LEE
- $n = 5$
- median: 0.21
- maximum: 0.38
- minimum: 0.13
- range: 0.25

2. CREST
- $n = 3$
- median: 0.67
- maximum: 0.85
- minimum: 0.61
- range: 0.24

3. STOSS
- $n = 12$
- median: 0.44
- maximum: 0.63
- minimum: 0.25
- range: 0.38

4. WESTERN HORN
- $n = 4$
- median: 0.34
- maximum: 0.38
- minimum: 0.30
- range: 0.08

5. EASTERN HORN
- $n = 5$
- median: 0.73
- maximum: 1.00
- minimum: 0.39
- range: 0.61

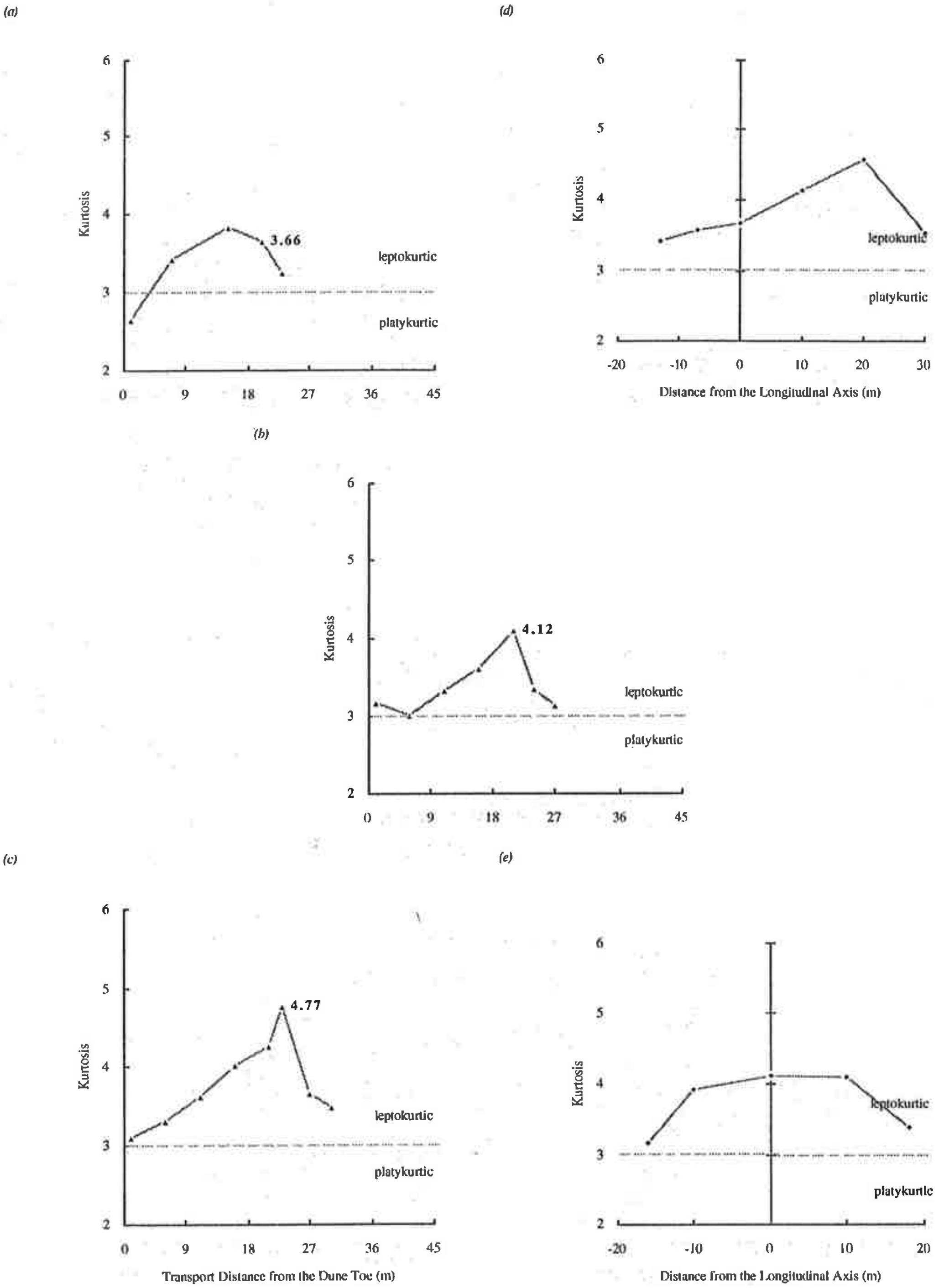

Figure 6.15 Longitudinal profiles *(a - c)* of kurtosis from the dune toe (lower stoss) towards dune lee, for dunes (3, 7, 8) and transverse profiles of dunes (3 & 7) in autumn 1993. (Bold numerals represent the kurtosis of the crest while negative *x* values represent the western horn, and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE FOURTH MOMENT (KURTOSIS)

1. LEE
- $n = 5$
- median: 3.35
- maximum: 3.66
- minimum: 3.14
- range: 0.52

2. CREST
- $n = 3$
- median: 4.12
- maximum: 4.77
- minimum: 3.66
- range: 1.11

3. STOSS
- $n = 12$
- median: 3.39
- maximum: 4.27
- minimum: 2.65
- range: 1.62

4. WESTERN HORN
- $n = 4$
- median: 3.50
- maximum: 3.92
- minimum: 3.16
- range: 0.76

5. EASTERN HORN
- $n = 5$
- median: 4.09
- maximum: 4.58
- minimum: 3.37
- range: 1.21

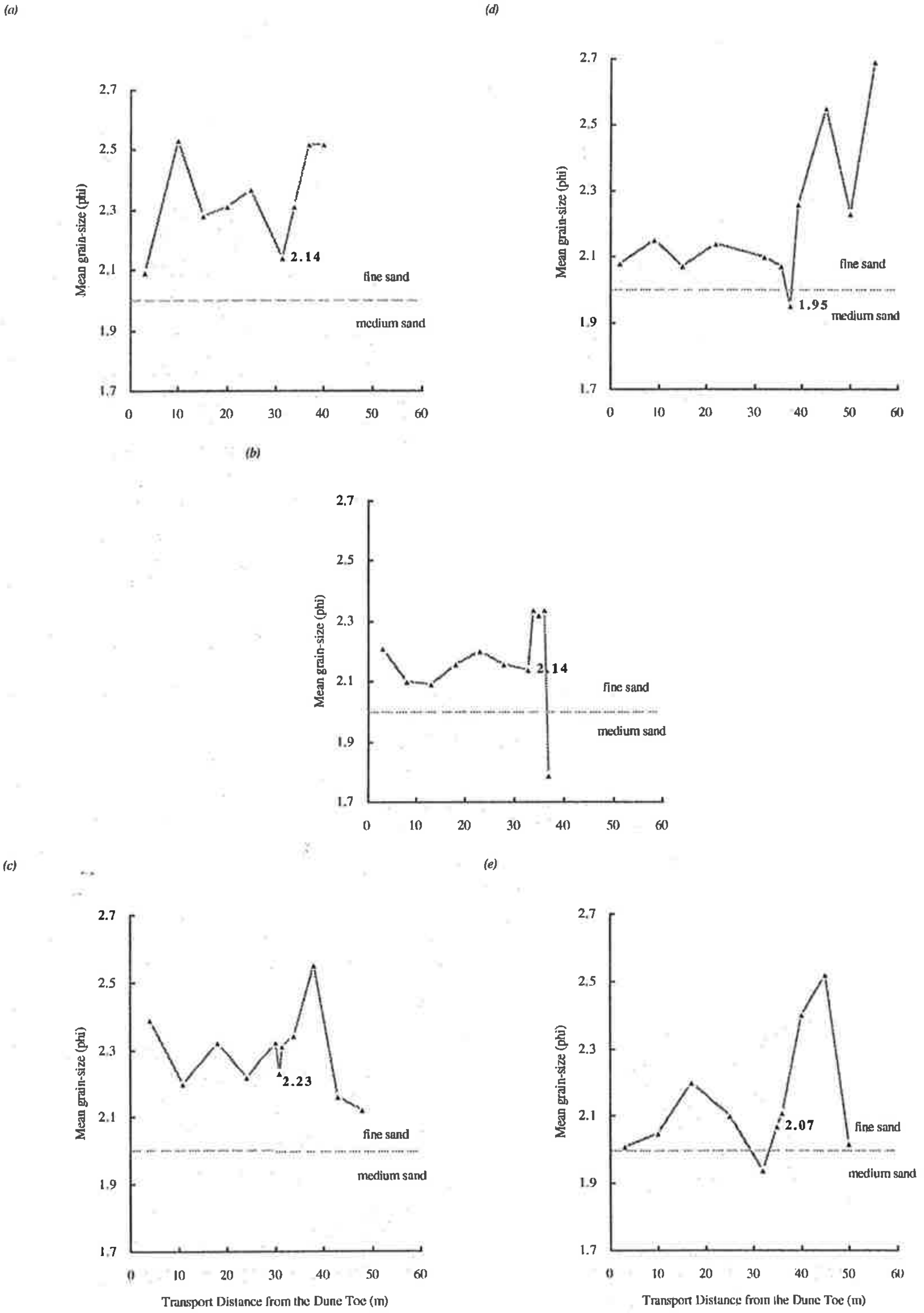

Figure 6.16 Longitudinal profiles of mean grain-size from the dune toe (lower stoss) towards the dune lee, for dunes (2, 7, 8, 9, 10) in spring 1994. (Bold numerals represent the mean grain-size of the crest).

DESCRIPTIVE STATISTICS FOR THE FIRST MOMENT (MEAN GRAIN-SIZE - ϕ units)

1. LEE
- $n = 19$
- median: 2.34
- maximum: 2.69
- minimum: 1.79
- range: 0.90

2. CREST
- $n = 13$
- median: 2.12
- maximum: 2.38
- minimum: 1.77
- range: 0.61

3. STOSS
- $n = 28$
- median: 2.16
- maximum: 2.53
- minimum: 1.94
- range: 0.59

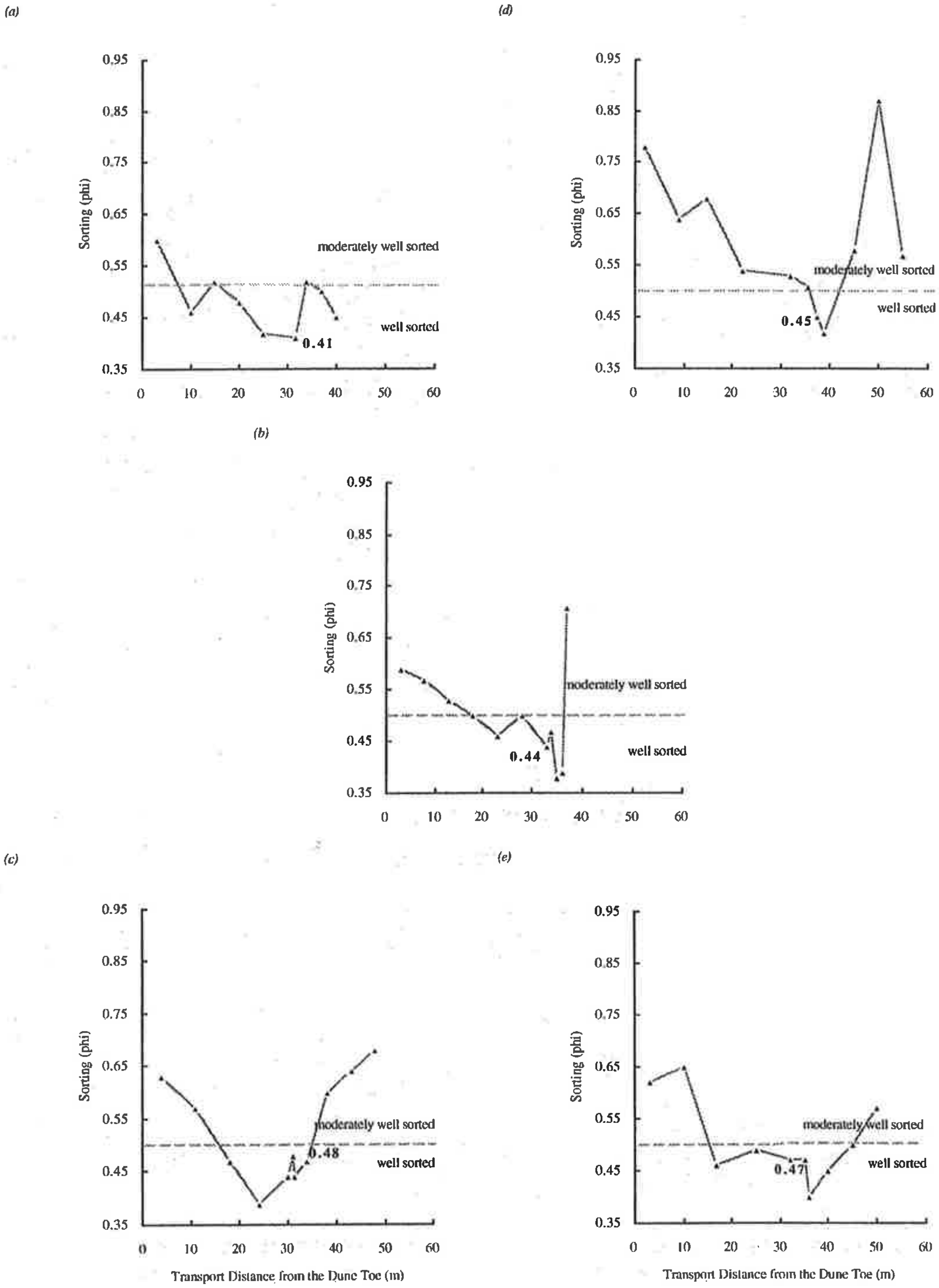

Figure 6.17 Longitudinal profiles of sorting from the dune toe (lower stoss) towards the dune lee, for dunes (2, 7, 8, 9, 10) in spring 1994. (Bold numerals represent the standard deviation (sorting) of the crest).

DESCRIPTIVE STATISTICS FOR THE SECOND MOMENT (SORTING)

1. LEE
- $n = 19$
- median: 0.50
- maximum: 0.87
- minimum: 0.38
- range: 0.49

2. CREST
- $n = 13$
- median: 0.45
- maximum: 0.53
- minimum: 0.41
- range: 0.12

3. STOSS
- $n = 28$
- median: 0.52
- maximum: 0.78
- minimum: 0.39
- range: 0.39

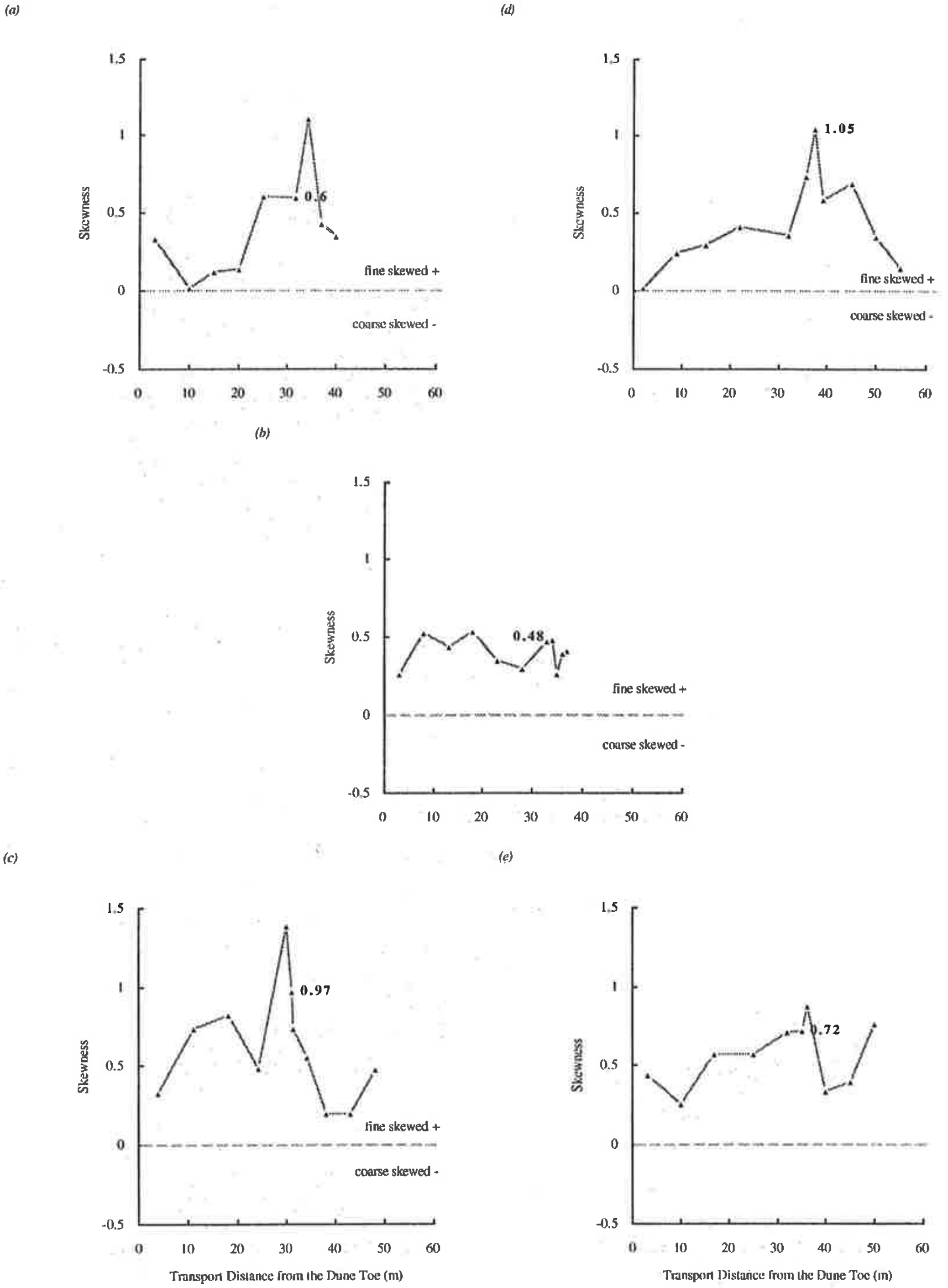

Figure 6.18 Longitudinal profiles of skewness for the sediment samples of dunes (2, 7, 8, 9, 10), from the dune toe (lower stoss) towards the dune lee in spri
(Bold numerals represent the third moment (skewness) of the crest).

DESCRIPTIVE STATISTICS FOR THE THIRD MOMENT (SKEWNESS)

1. LEE
- n = 19
- median: 0.42
- maximum: 1.11
- minimum: 0.15
- range: 0.96

2. CREST
- n = 13
- median: 0.74
- maximum: 1.08
- minimum: 0.44
- range: 0.64

3. STOSS
- n = 28
- median: 0.43
- maximum: 1.39
- minimum: 0.02
- range: 1.37

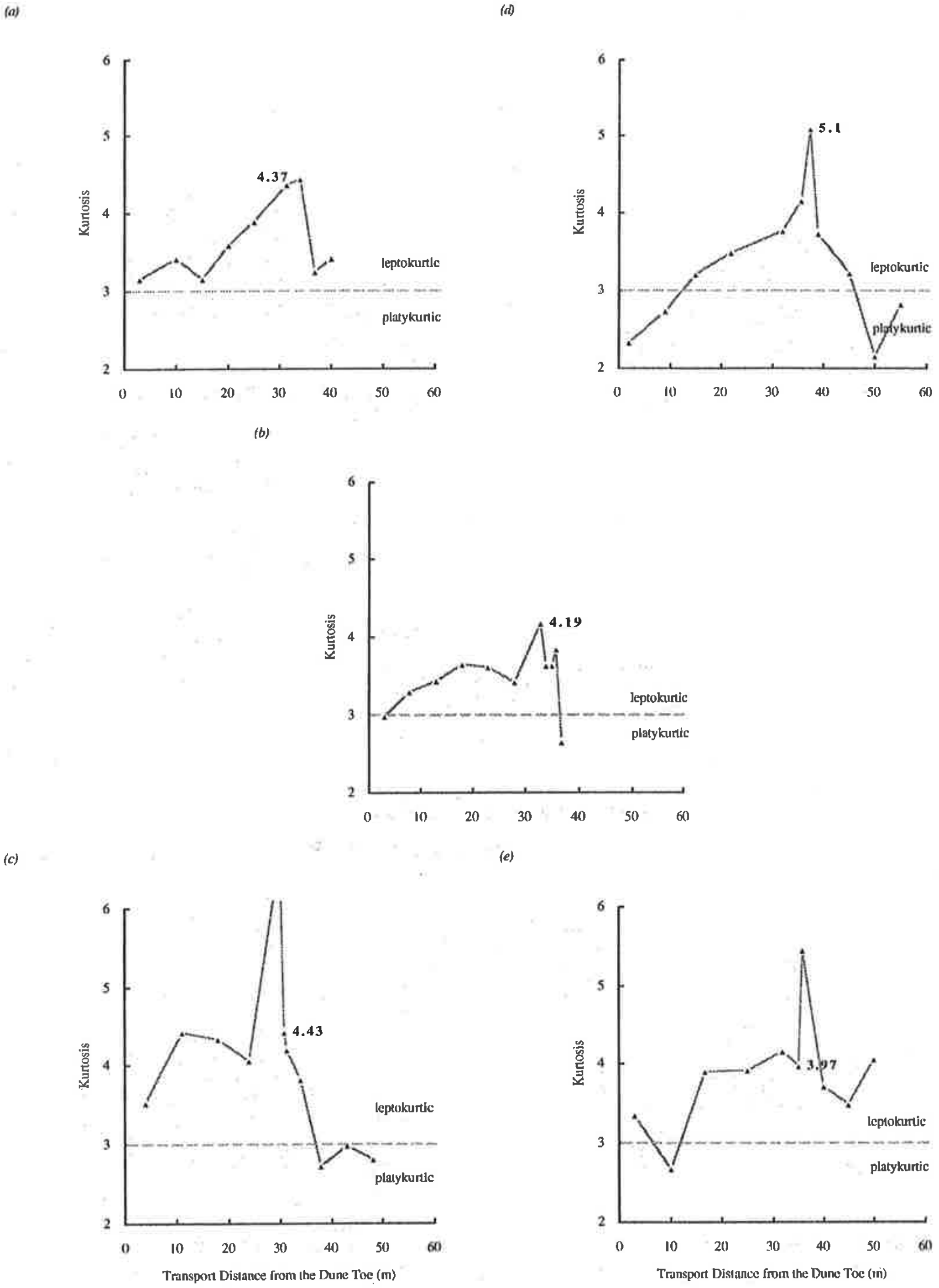

Figure 6.19 Longitudinal profiles of kurtosis for the sediment samples of dunes (2, 7, 8, 9, 10), from the dune toe (lower stoss) towards the dune lee in spring (**Bold numerals represent the fourth moment (kurtosis) of the crest**).

DESCRIPTIVE STATISTICS FOR THE FOURTH MOMENT (KURTOSIS)

1. LEE

- *n* = 19
- median: 3.49
- maximum: 5.40
- minimum: 2.17
- range: 3.28

2. CREST

- *n* = 13
- median: 4.21
- maximum: 5.10
- minimum: 3.78
- range: 1.32

3. STOSS

- *n* = 28
- median: 3.57
- maximum: 6.67
- minimum: 2.33
- range: 4.34

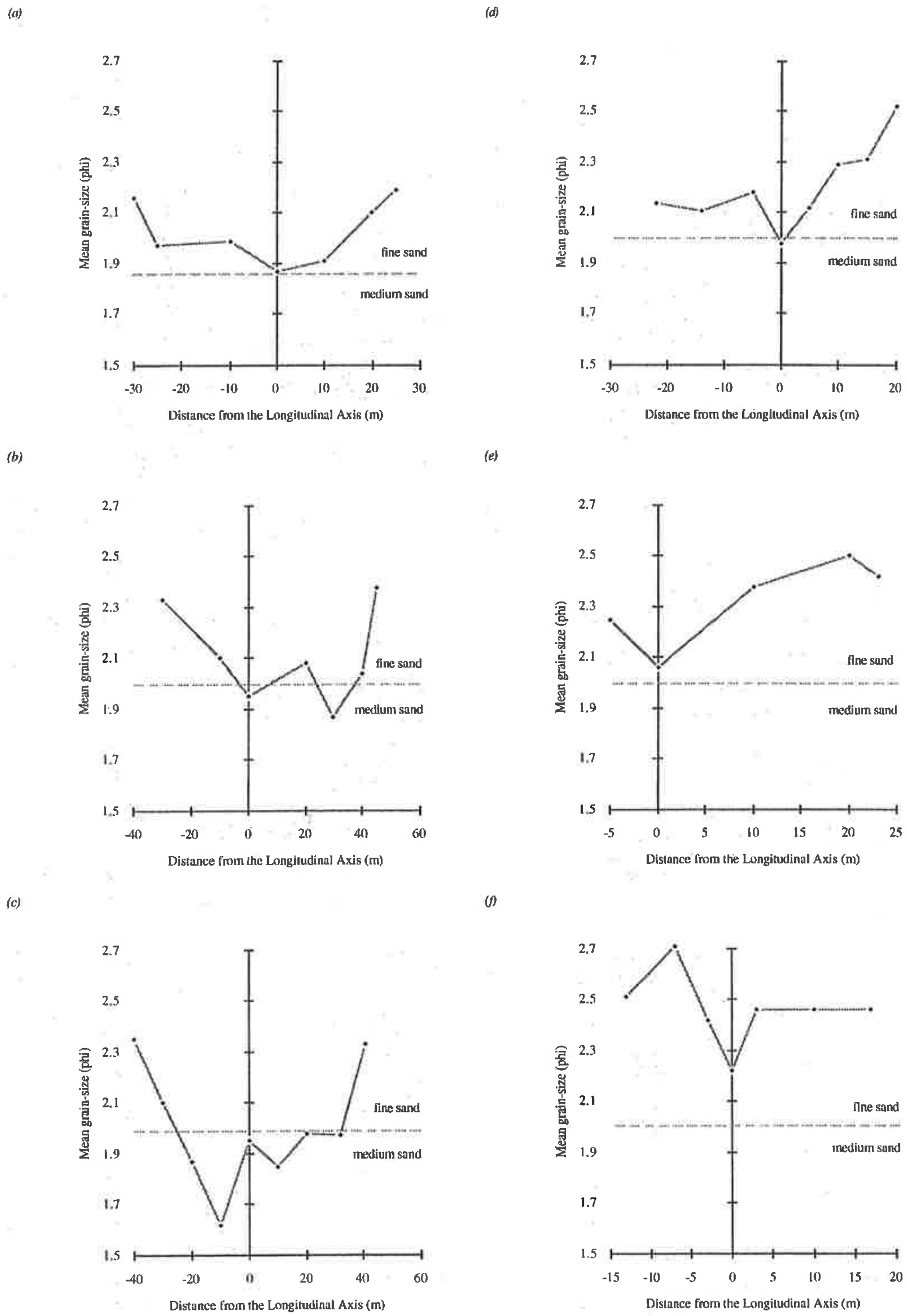

Figure 6.20 Transverse profiles of mean grain-size for dunes (1 - 6) in winter 1991. (Negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE FIRST MOMENT (MEAN GRAIN-SIZE - ϕ units)

1. WESTERN HORN

- $n = 16$
- median: 2.15
- maximum: 2.71
- minimum: 1.62
- range: 1.09

2. EASTERN HORN

- $n = 21$
- median: 2.29
- maximum: 2.52
- minimum: 1.85
- range: 0.67

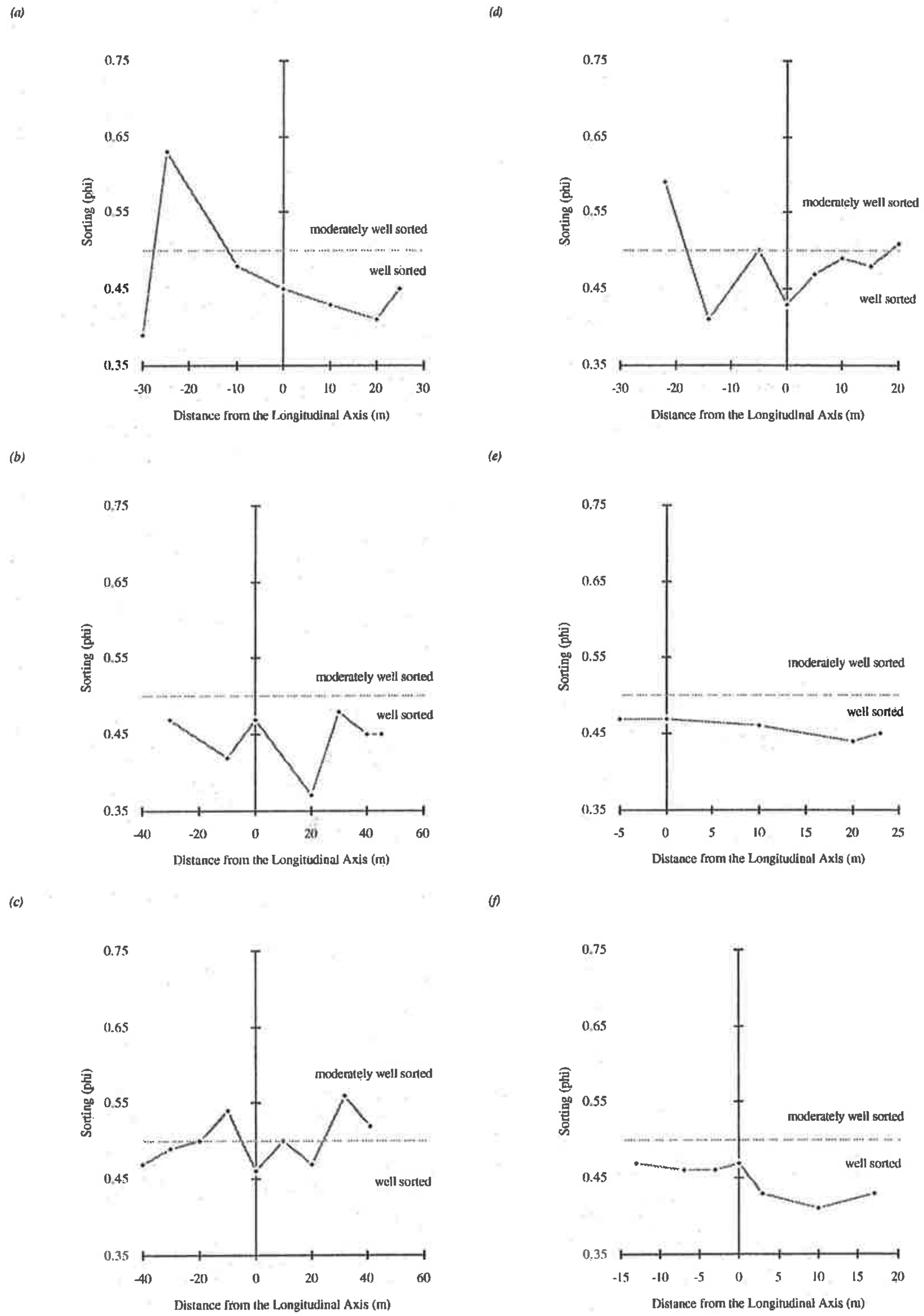

Figure 6.21 Transverse profiles of sorting for dunes (1 - 6) in winter 1991. (Negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE SECOND MOMENT (SORTING)

1. WESTERN HORN

- *n* = 16
- median: 0.47
- maximum: 0.63
- minimum: 0.39
- range: 0.24

2. EASTERN HORN

- *n* = 21
- median: 0.45
- maximum: 0.56
- minimum: 0.37
- range: 0.19

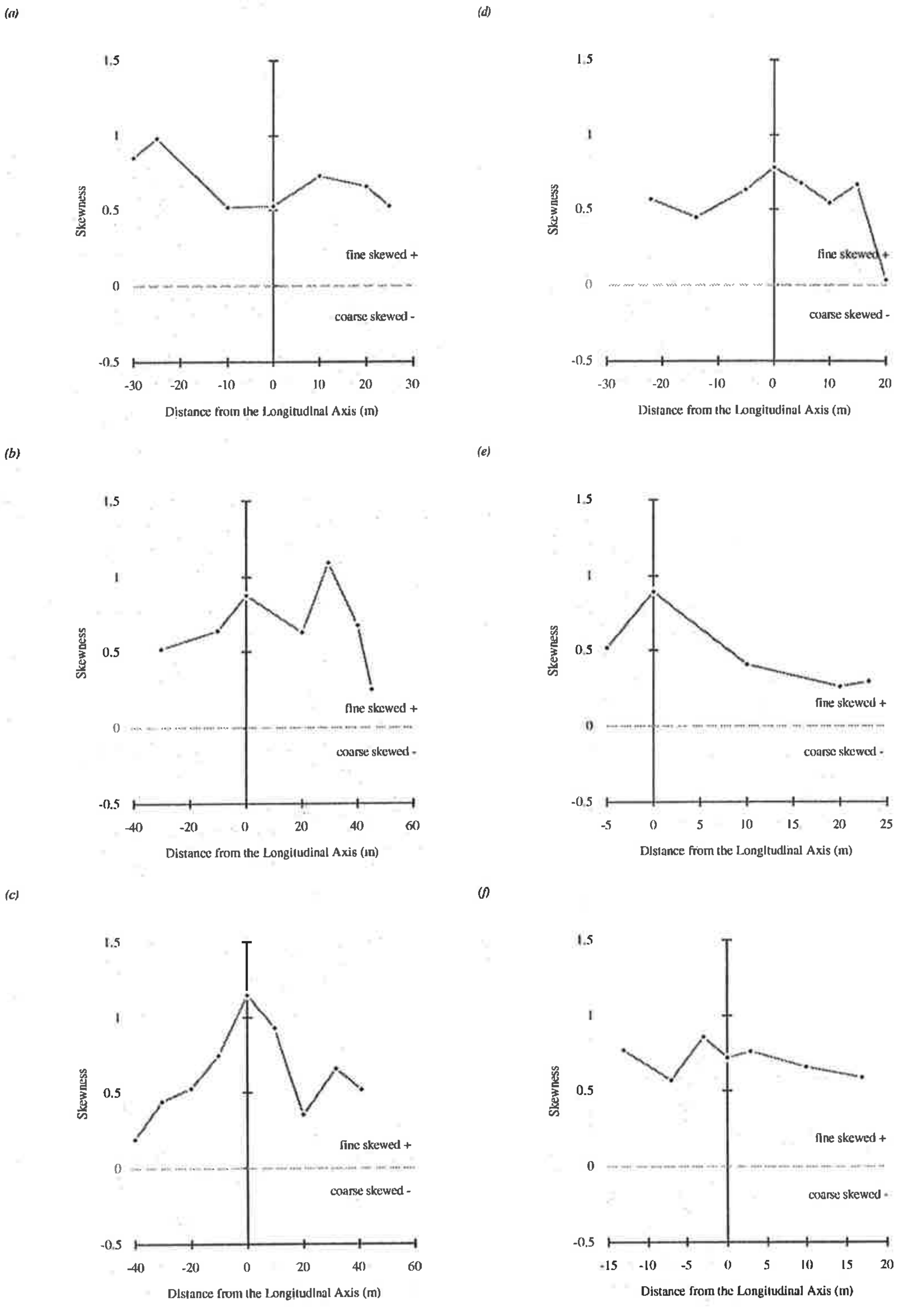

Figure 6.22 Transverse profiles of skewness for dunes (1 - 6) in winter 1991. (Negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE THIRD MOMENT (SKEWNESS)

1. WESTERN HORN
- *n* = 16
- median: 0.57
- maximum: 0.98
- minimum: 0.19
- range: 0.79

2. EASTERN HORN
- *n* = 21
- median: 0.63
- maximum: 1.09
- minimum: 0.04
- range: 1.05

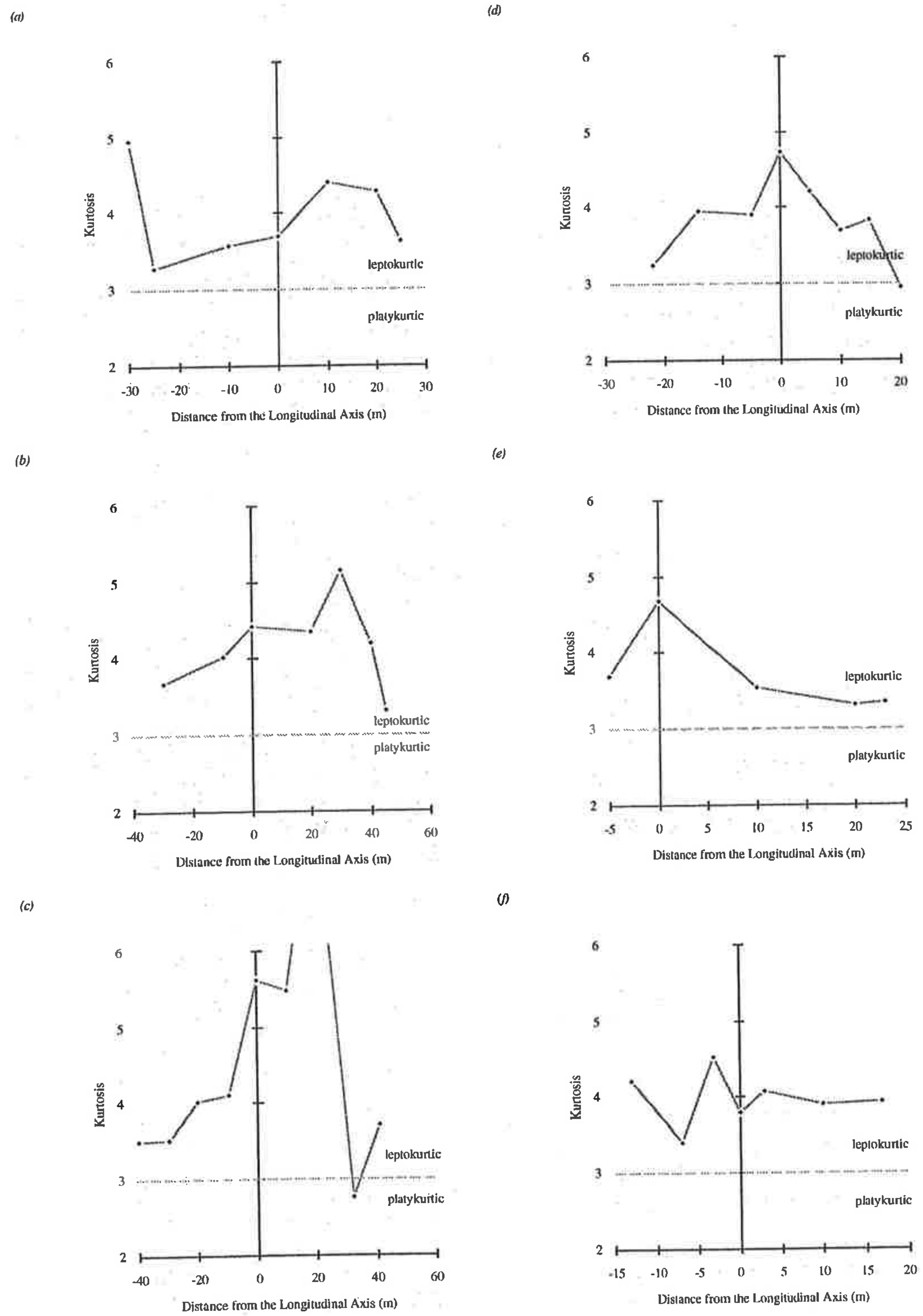

Figure 6.23 Transverse profiles of kurtosis for dunes (1 - 6) in winter 1991. (Negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE FOURTH MOMENT (KURTOSIS)

1. WESTERN HORN

- $n = 16$
- median: 3.81
- maximum: 4.96
- minimum: 3.25
- range: 1.71

2. EASTERN HORN

- $n = 21$
- median: 3.93
- maximum: 8.10
- minimum: 2.95
- range: 5.15

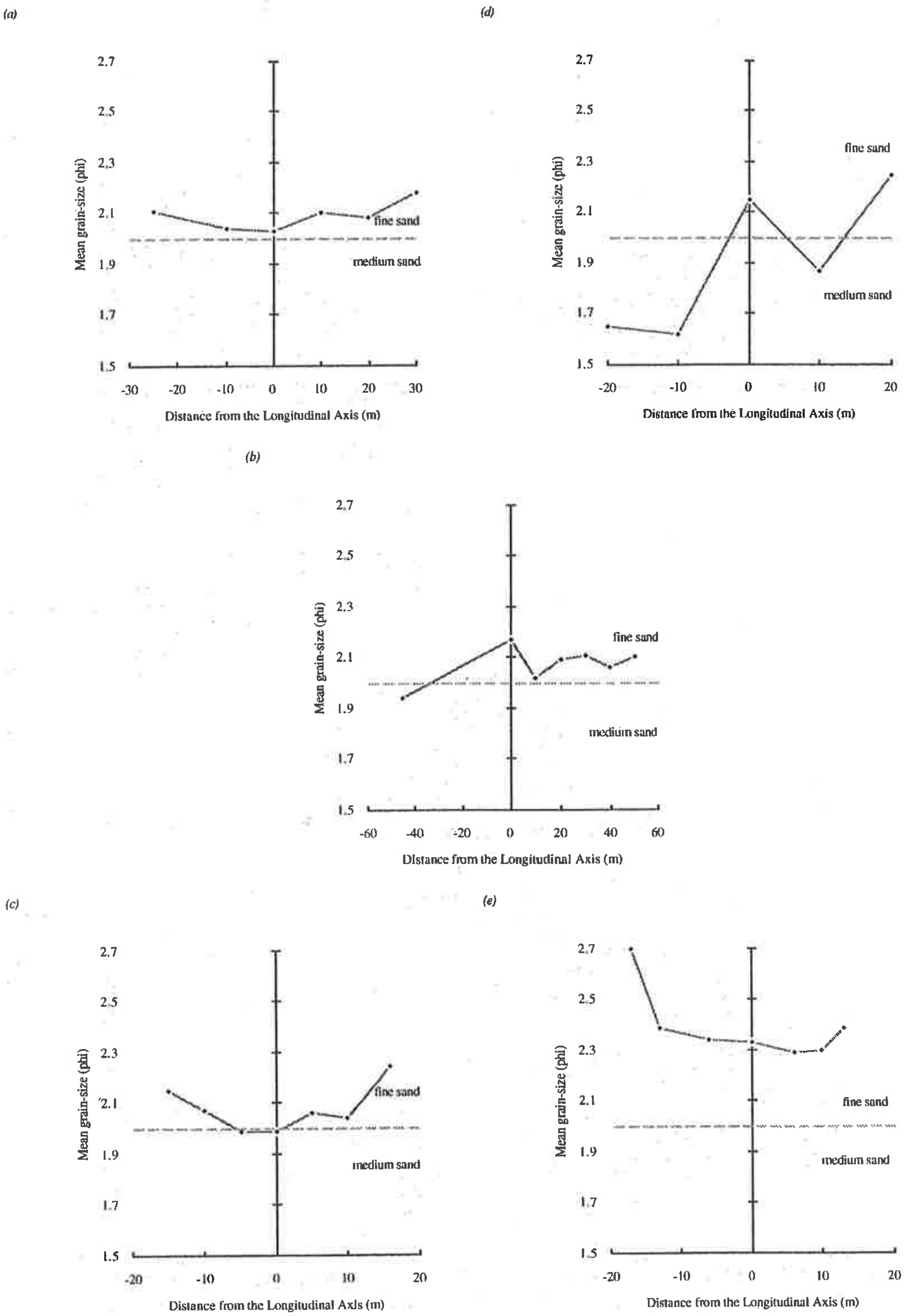

Figure 6.24 Transverse profiles of mean grain-size for dunes (1, 2, 3a, 3b, 6) in summer 1992. (Negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE FIRST MOMENT (MEAN GRAIN-SIZE)

1. WESTERN HORN
- $n = 11$
- median: 2.07
- maximum: 2.70
- minimum: 1.62
- range: 1.08

2. EASTERN HORN
- $n = 16$
- median: 2.10
- maximum: 2.39
- minimum: 1.87
- range: 0.52

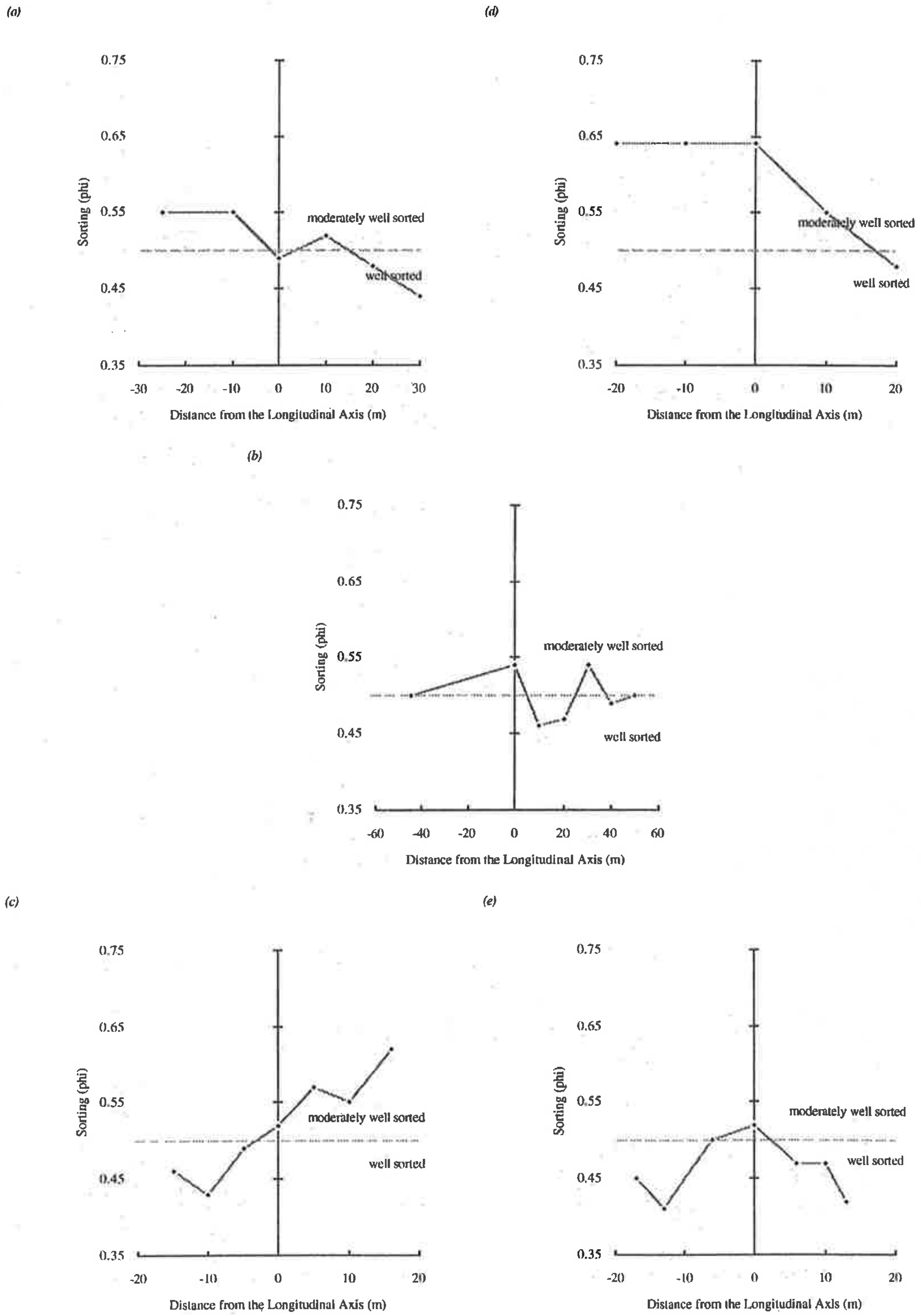

Figure 6.25 Transverse profiles of sorting for dunes (1, 2, 3a, 3b, 6) in summer 1992. (Negative *x* values represent the western horn while positve values portray the eastern horn).

DESCRIPTIVE STATISTICS FOR THE SECOND MOMENT (SORTING)

1. WESTERN HORN
- $n = 11$
- median: 0.50
- maximum: 0.64
- minimum: 0.41
- range: 0.23

2. EASTERN HORN
- $n = 16$
- median: 0.49
- maximum: 0.62
- minimum: 0.42
- range: 0.20

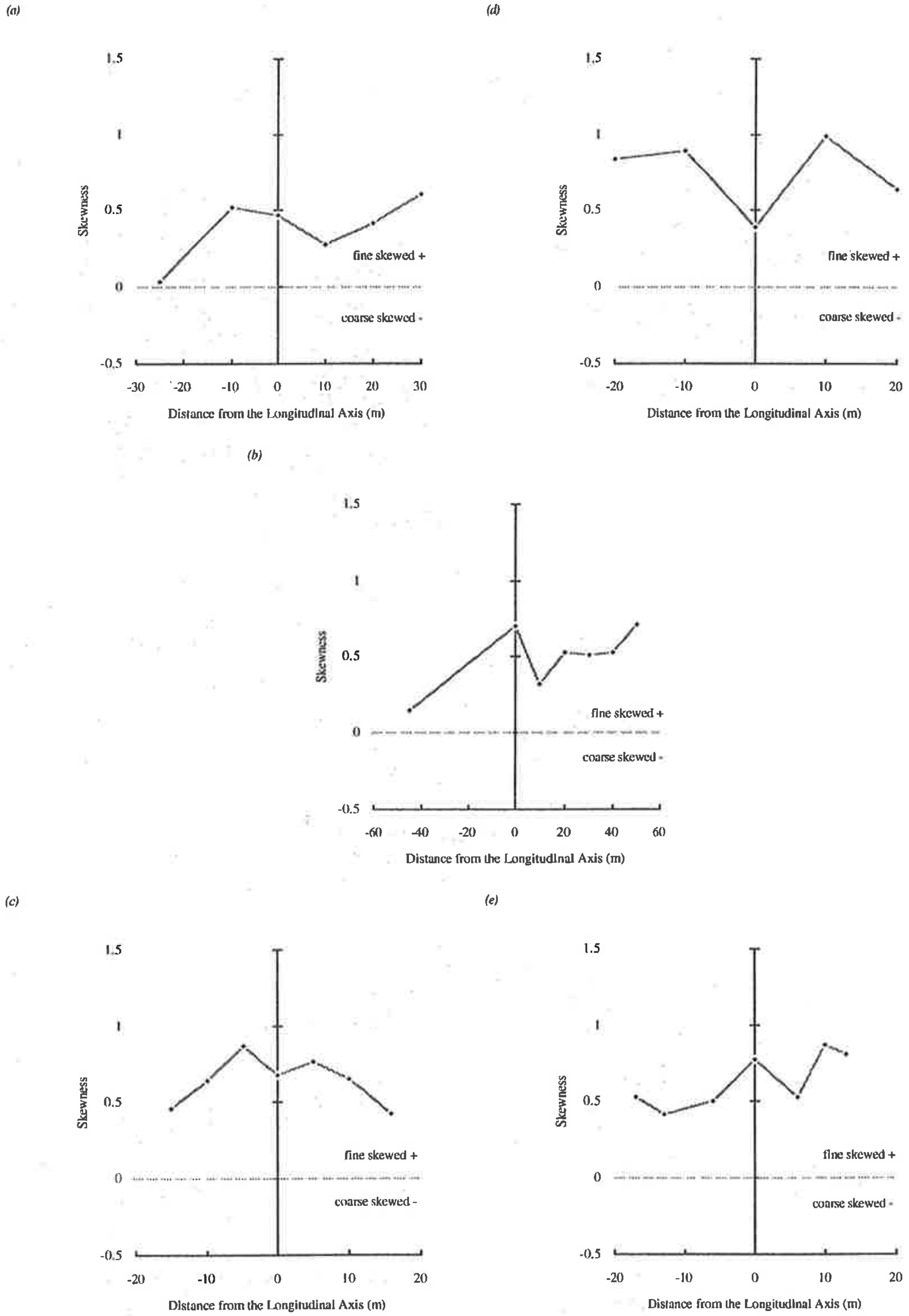

Figure 6.26 Transverse profiles of skewness for dunes (1, 2, 3a, 3b, 6) in summer 1992. (Negative *x* values represent the western horn while positve values portray the eastern horn).

DESCRIPTIVE STATISTICS FOR THE THIRD MOMENT (SKEWNESS)

1. WESTERN HORN
- *n* = 11
- median: 0.52
- maximum: 0.89
- minimum: 0.04
- range: 0.85

2. EASTERN HORN
- *n* = 16
- median: 0.57
- maximum: 0.99
- minimum: 0.28
- range: 0.71

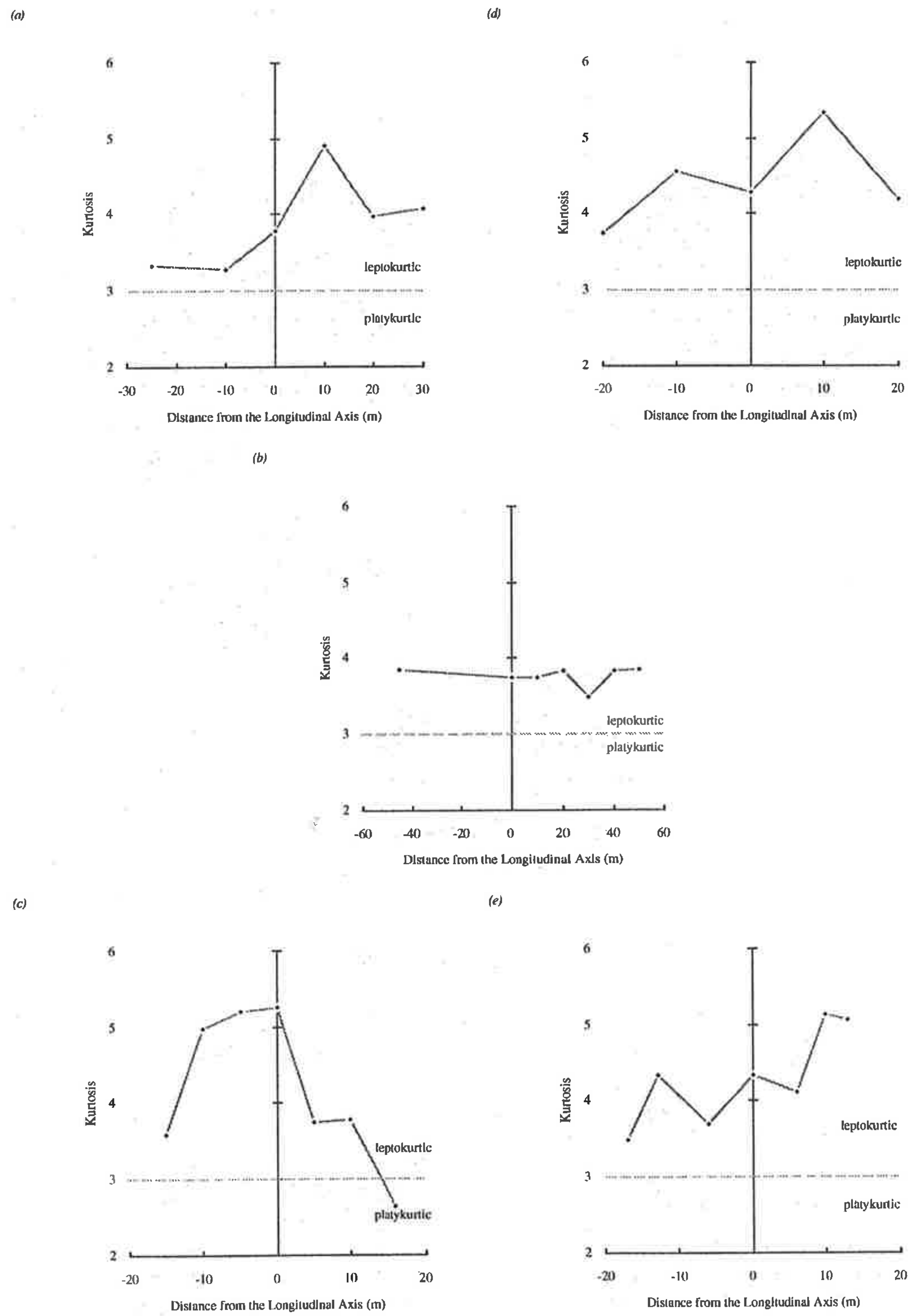

Figure 6.27 Transverse profiles of kurtosis for dunes (1, 2, 3a, 3b, 6) in summer 1992. (Negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE FOURTH MOMENT (KURTOSIS)

1. WESTERN HORN

- *n* = 11
- median: 3.75
- maximum: 5.20
- minimum: 3.28
- range: 1.92

2. EASTERN HORN

- *n* = 16
- median: 3.92
- maximum: 5.34
- minimum: 2.64
- range: 2.70

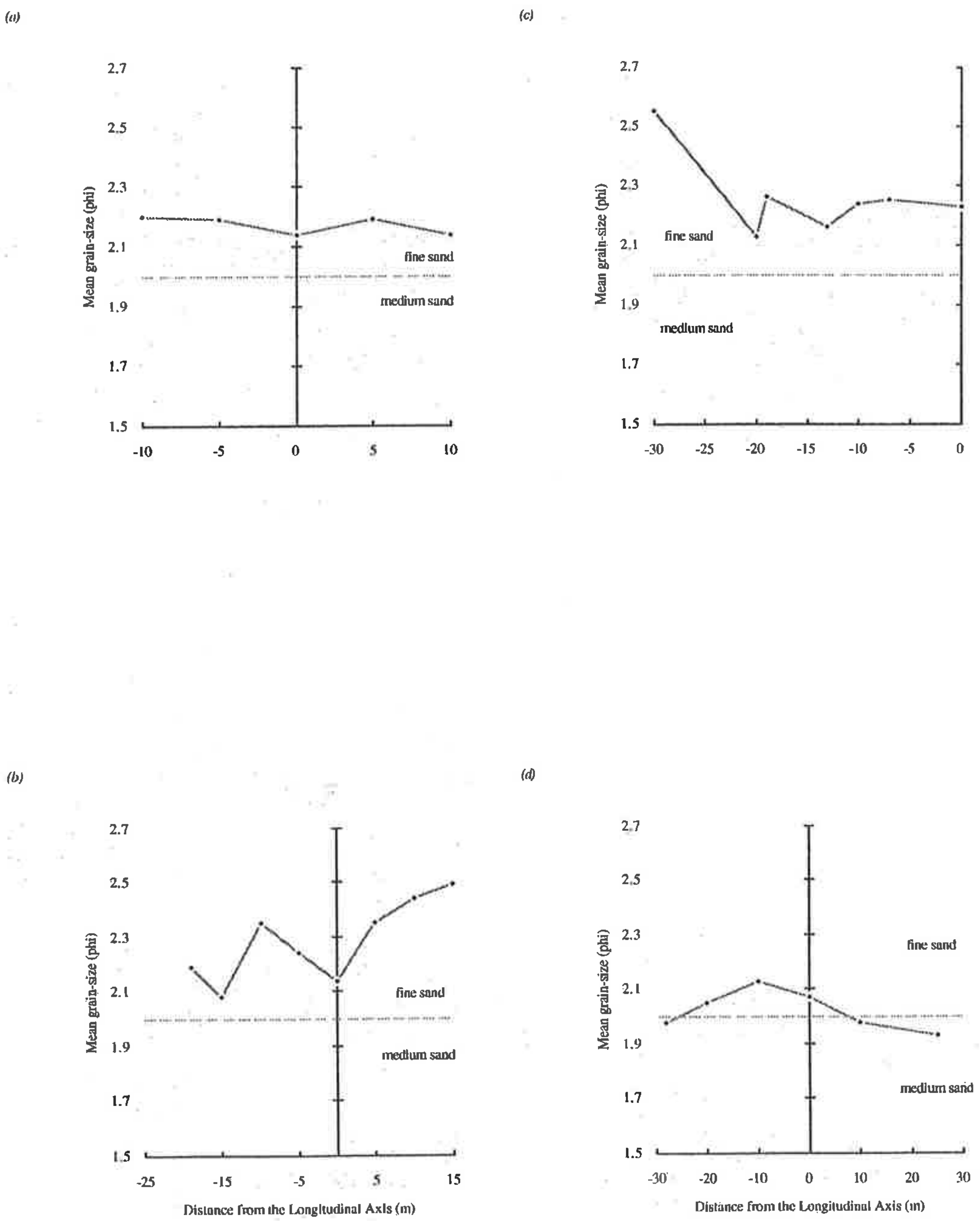

Figure 6.28 Transverse profiles of mean grain-size for dunes (2, 7, 8, 10) in spring 1994. (Negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE FIRST MOMENT (MEAN GRAIN-SIZE - ϕ units)

1. WESTERN HORN

- $n = 15$
- median: 2.19
- maximum: 2.55
- minimum: 1.98
- range: 0.57

2. EASTERN HORN

- $n = 7$
- median: 2.19
- maximum: 2.49
- minimum: 1.93
- range: 0.56

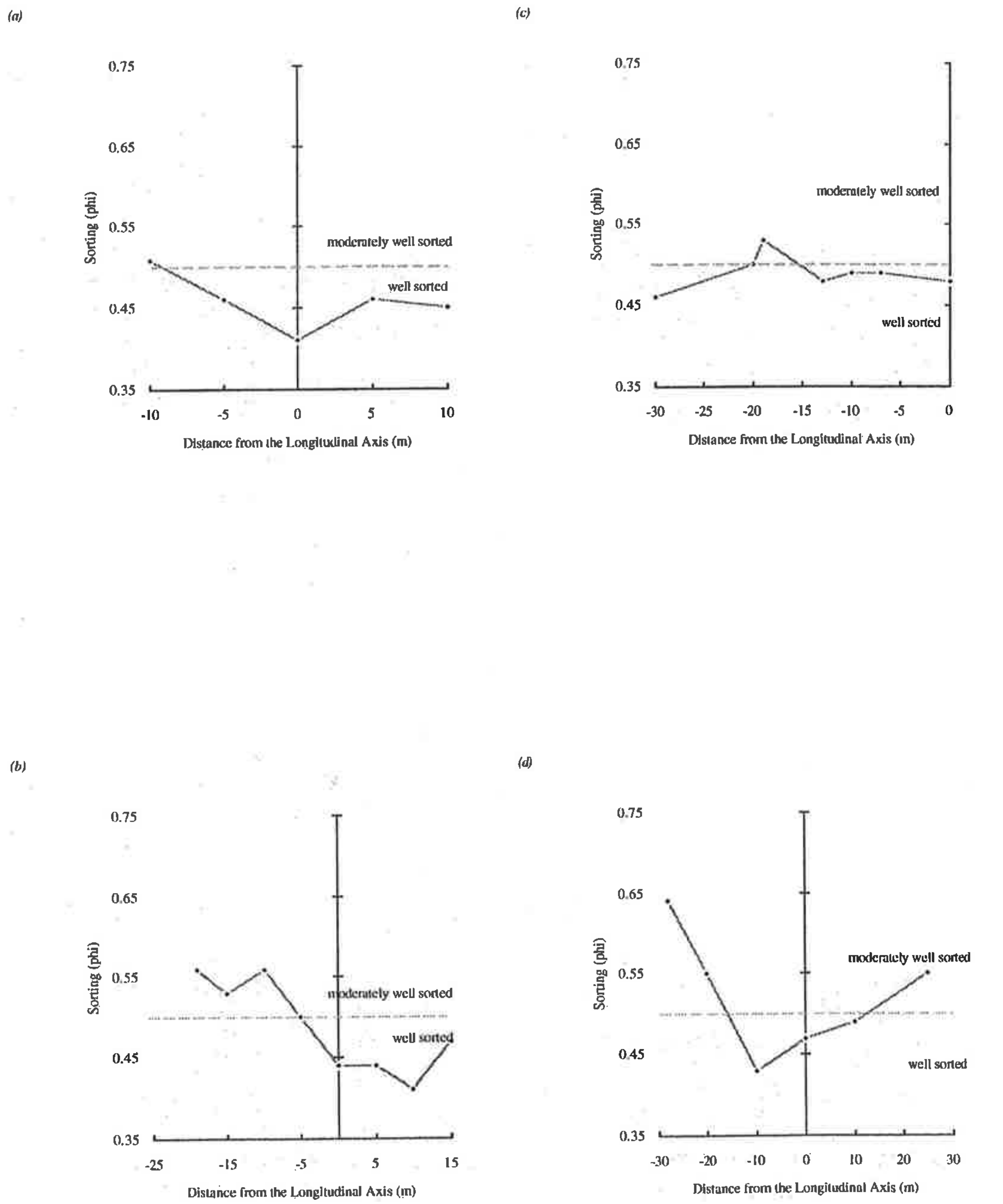

Figure 6.29 Transverse profiles of sorting for dunes (2, 7, 8, 10) in spring 1994. (Negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE SECOND MOMENT (SORTING)

1. WESTERN HORN
- *n* = 15
- median: 0.50
- maximum: 0.64
- minimum: 0.43
- range: 0.21

2. EASTERN HORN
- *n* = 7
- median: 0.46
- maximum: 0.55
- minimum: 0.41
- range: 0.14

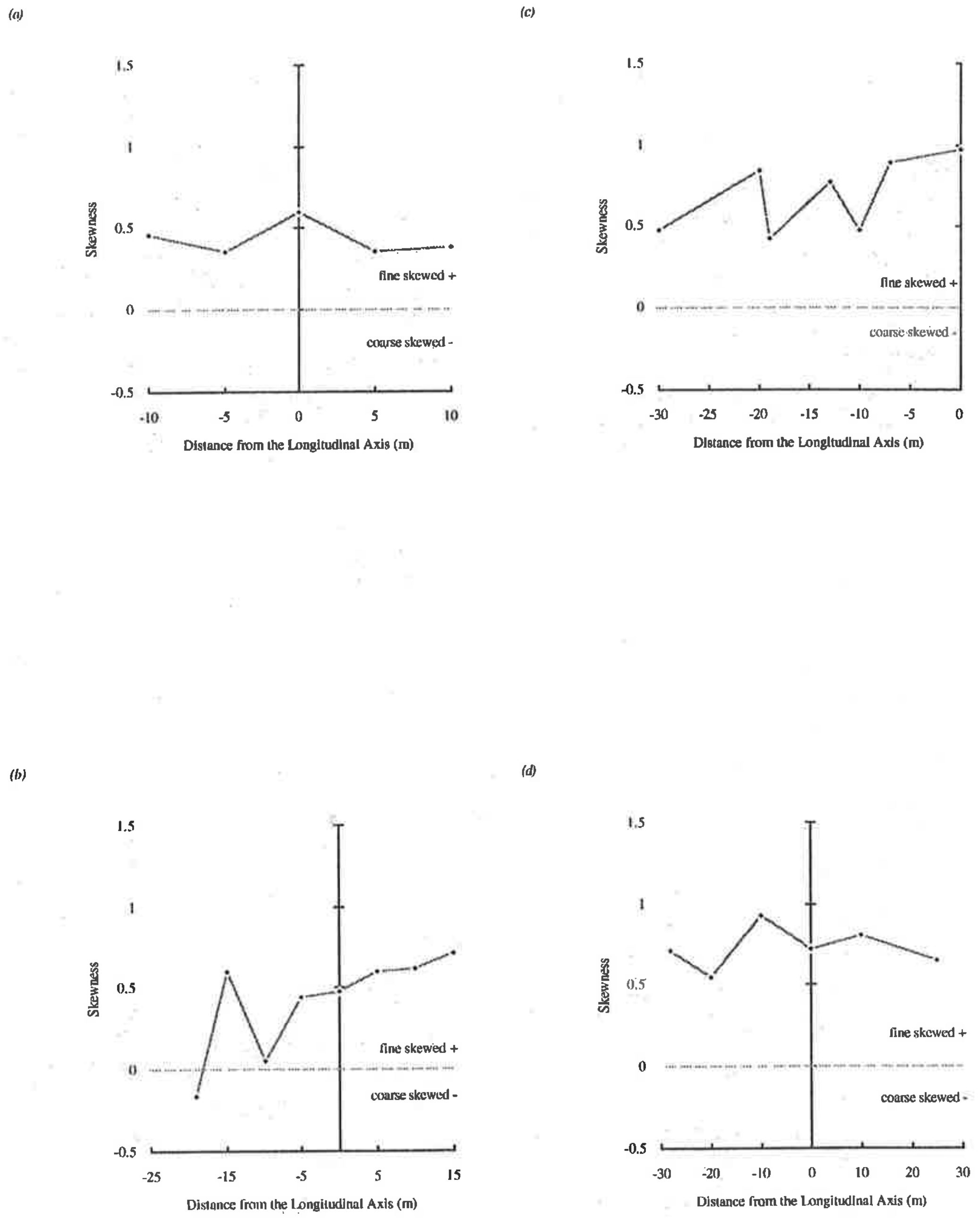

Figure 6.30 Transverse profiles of skewness for dunes (2, 7, 8, 10) spring 1994. (Negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE THIRD MOMENT (SKEWNESS)

1. WESTERN HORN
- $n = 15$
- median: 0.48
- maximum: 0.93
- minimum: -0.16
- range: 1.09

2. EASTERN HORN
- $n = 7$
- median: 0.62
- maximum: 0.81
- minimum: 0.36
- range: 0.45

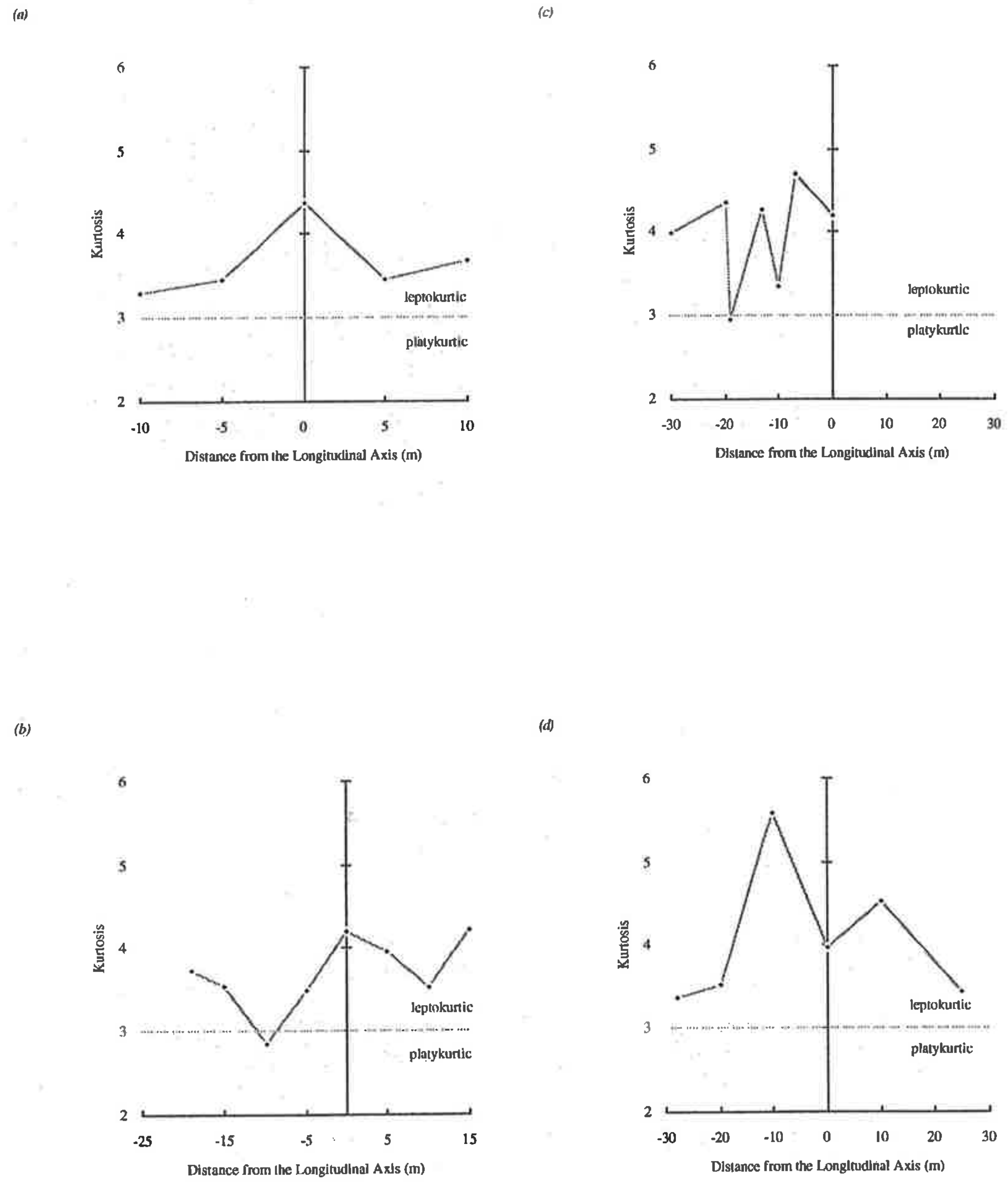

Figure 6.31 Transverse profiles of kurtosis for dunes (2, 7, 8, 10) in spring 1994. (Negative *x* values represent the western horn and positive values the eastern horn).

DESCRIPTIVE STATISTICS FOR THE FOURTH MOMENT (KURTOSIS)

1. WESTERN HORN
- $n = 15$
- median: 3.53
- maximum: 5.59
- minimum: 2.84
- range: 2.75

2. EASTERN HORN
- $n = 7$
- median: 3.68
- maximum: 4.53
- minimum: 3.43
- range: 1.13

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