Pergamon

Journal of African Earth Sciences, Vol. 29, No. 4, pp. 629–642, 1999

PII:S0899-5362(99)00120-7

Tertiary Tsondab Sandstone Formation: preliminary bedform reconstruction and comparison to modern Namib Sand Sea dunes

GARY KOCUREK,[1,*] NICHOLAS LANCASTER,[2] MARY CARR[1,3] and ANDY FRANK[1,4]

[1]Department of Geological Sciences, University of Texas, Austin, Texas 78712 USA

[2]Desert Research Institute, Quaternary Sciences Center, PO Box 60220, Reno, Nevada 89506 USA

[3]Current address: Department of Geology and Geological Engineering, Colorado School of Mines, Golden, Colorado 80401 USA

[4]Current address: Chemeketa Community College, PO Box 14007, Salem, Oregon 97309 USA

ABSTRACT—The Tertiary Tsondab Sandstone Formation, which underlies much of the present Namib Sand Sea, is a key element in understanding the Cenozoic evolution of the Namib Desert. Outcrops of the aeolian facies of the Tsondab Sandstone at Elim and Diep Rivier consist of two sequences of bioturbated cross-strata separated by likely formation-scale surfaces of stabilisation. Cross-strata consist of scalloped sets about 200 m in width and separated by southeast dipping bounding surfaces. Internally, sets contain reactivation surfaces of probable seasonal origin. The north to south-southeast dipping foresets define crescent shapes with a trough axis trending northeast. Although additional data are needed to define the Tsondab bedform, the outcrop data is best satisfied in computer simulations by north trending, east migrating main bedforms, which had relatively large and slow-moving dunes superimposed upon their eastern flanks and migrated to the north. Foresets dipping to the south to south-southwest at Elim suggest that superimposed dunes also occurred on the western flanks of the main bedform and migrated to the south, but that their record was largely lost with net eastward migration of the main bedform. This preliminary Tsondab model shares attributes such as trend, scale of cross-strata, and presence of scalloped sets with reactivation surfaces with computer models of the modern linear dunes in which large-scale sinuosity migrates alongcrest to the north. Differences emerge in the overall set architecture and the orientation of cross-strata and bounding surfaces, as well as the degree of vegetation that must have characterised Tsondab dunes. © 2000 Elsevier Science Limited. All rights reserved.

RÉSUMÉ—La Formation Tertiaire des Grès de Tsondab, sous-jacente à l'essentiel de la Mer de Sable actuelle du Namib, est un élément-clé pour comprendre l'évolution cénozoïque du Désert de Namib. Les affleurements du faciès éolien des Grès de Tsondab à Elim et Diep Rivier forment deux séquences de strates obliques avec bioturbation séparées par des surfaces de stabilisation vraisemblablement à l'échelle de la formation. Les strates à stratification oblique forment des couches festonnées d'environ 200 m de large et séparés par des surfaces à pendage sud-est. A l'intérieur, les couches montrent des surfaces de réactivation d'origine probablement saisonnière. Les pentes frontales de pendanges nord à sud-sud-est définissent des croissants dont l'axe de la concavité a une orientation NE. Bien que d'autres données soient nécessaire pour définir la forme du lit de Tsondab, nos observations de terrain sont compatibles avec une simulation par ordinateur utilisant un lit majeur de direction nord et migrant vers l'est, contenant des dunes superposées sur son flanc oriental relativement grandes et se déplaçant lentement vers le nord. A Elim, les pentes frontales de pendage sud à sud-sud-ouest suggèrent que les dunes superposées existaient aussi

* Corresponding author
garyk@mail.utexas.edu (G. Kocurek)

sur les flancs occidentaux du lit majeur et migraient vers le sud mais leur enregistrement sédimentaire a été largement perdu lorsque le réseau du lit majeur a migré vers l'est. Ce modèle préliminaire de Tsondab présente des caractères, comme la direction, l'échelle des strates obliques, la présence de couches festonnées avec surfaces de réactivation, en commun avec les modèles informatiques de dunes linéaires modernes où la sinuosité de grande longueur d'onde migre vers le nord. Des différences apparaissent dans l'architecture d'ensemble et l'orientation des strates obliques et des surfaces de stabilisation, ainsi que dans le taux de végétation qui a pu caractériser les dunes de Tsondab.

(Received 28/9/98: revised version received 29/6/99: accepted 29/6/99)

INTRODUCTION

The Namib Desert, lying west of the Great Escarpment between the Carunjamba River in Angola and the Olifants River in South Africa, houses the Namib Sand Sea, which covers about 34,000 km^2 of Namibia south of the Kuiseb River. The antiquity of this region as a desert and the controls on the regional climate, physiographic evolution, and sediment history have long been debated (see reviews in Ward *et al.*, 1983; Ward and Corbett, 1990). One key element in the Namibian geological history is the Tsondab Sandstone, which underlies much of the present Namib Sand Sea between Luederitz and the Kuiseb River (Fig. 1). The Tsondab Sandstone has been taken as representing a proto-Namib sand sea (e.g. Ward *et al.*, 1983; Ward, 1987; Ward and Corbett, 1990), in which its age and origin could provide direct evidence for the antiquity of desert conditions and form a basis for comparison to the modern sand sea in terms of dune type and wind regime.

As a stratigraphical unit, the Tsondab Sandstone Formation is bounded below by the Namib Unconformity Surface, an erosional surface developed upon Late Precambrian bedrock and forming the fundamental datum for Cenozoic strata of the region (see Ward, 1987). Within the proto-Kuiseb and -Gaub Valleys, the Tsondab Sandstone is overlain by the Karpenkliff Conglomerate or, where this unit is absent, the yet younger Kamberg Calcrete. The Tsondab is believed to reach up to 220 m in thickness (Barnard, 1973), but outcrops are less than 100 m in thickness and the unit is in general poorly exposed.

The Tsondab Sandstone is widely accepted as largely aeolian in origin (Martin, 1950; Barnard, 1973; Ollier, 1977; Besler and Marker, 1979; Ward, 1987; Ward and Corbett, 1990). In the most comprehensive treatments of the Tsondab Sandstone, Ward (1987, pp. 10-15; 1988) recognises six facies, ranging from basal breccias and conglomerates, to cross-stratified to massive sandstones that comprise most of the unit, to local carbonate units. As with many aeolian sandstones, the Tsondab has proven difficult to date (Ward, 1987, pp. 14-15; Ward and Corbett, 1990, pp. 21-23), with the best resolution now derived from a biostratigraphy based on gigantic avian eggshells (Pickford *et al.*, 1995). This biostratigraphy argues that Tsondab and associated sandstones accumulated from probably the pre-Miocene to the Quaternary, with the bulk of the Tsondab having formed during the Miocene.

Although the avian biostratigraphy suggests that the Tsondab as a stratigraphical unit is complex and that aeolian conditions occurred in the region before the Miocene, comparisons to the modern sand sea are not well-documented. Previous works (Barnard, 1973; Besler and Marker, 1979) report northwest to northeast dip directions for the foresets, suggesting to Ward (1987, 1988) and Ward and Corbett (1990) dominant winds from the south as in the modern sand sea, but the detailed reconstruction

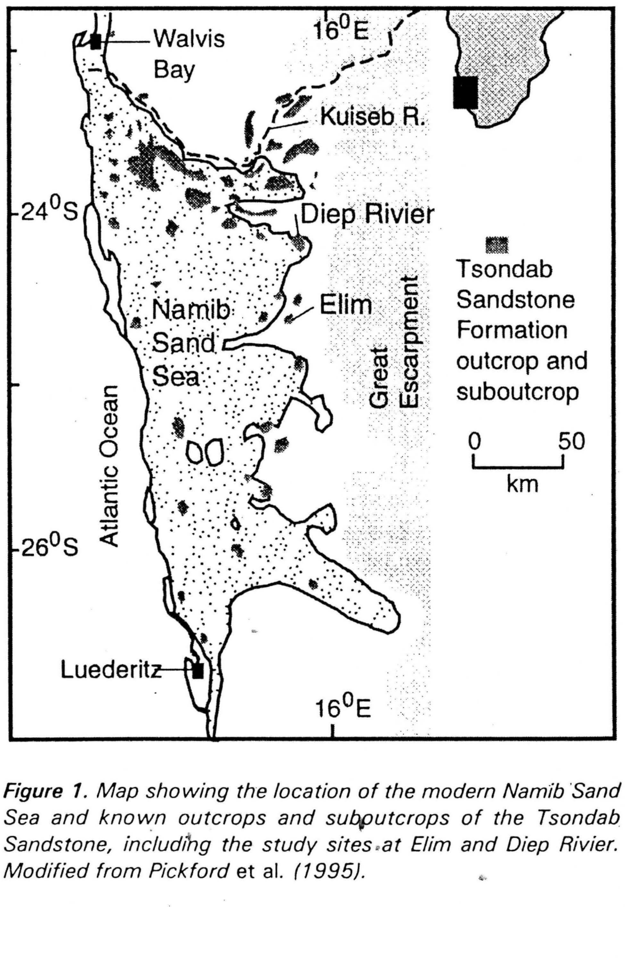

Figure 1. Map showing the location of the modern Namib Sand Sea and known outcrops and suboutcrops of the Tsondab Sandstone, including the study sites at Elim and Diep Rivier. Modified from Pickford et al. *(1995).*

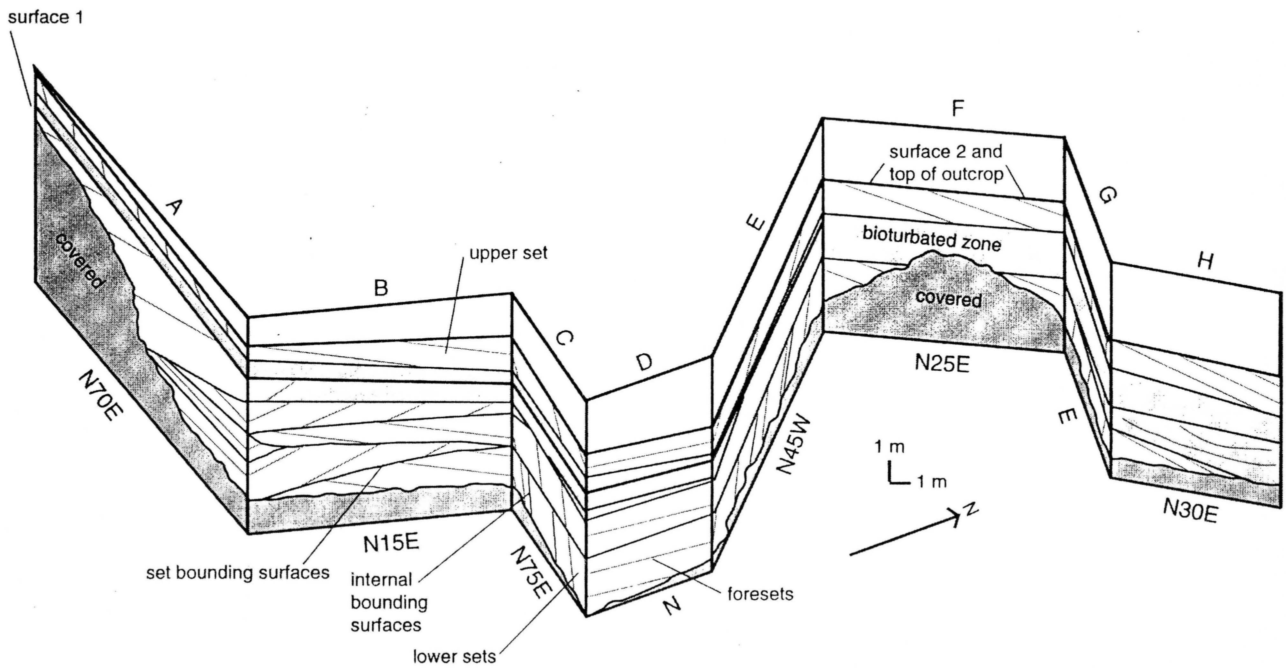

Figure 2. Scaled drawing of the outcrop at Elim, indicated on Fig. 1 and discussed in the text. This diagram is based upon field measurements and drawings, and a photo mosaic. Foreset and bounding surface orientations were measured where possible in the field, and are generalised here and shown with their apparent dip directions for each wall orientation.

of the Tsondab dunes needed to form a sound comparison does not exist. Although the morphology of the modern dunes can be seen, and limited trenching and natural exposures of the internal structure occur (McKee, 1982; Ward, 1987, 1988), the large-scale internal structure of the cross-strata is unknown.

Two of the better exposures of the Tsondab Sandstone, at Elim and Diep Rivier (Fig. 1), were studied with the singular goal of reconstructing the bedforms that gave rise to the cross-strata. The morphology of the modern dunes of the Namib and the existing data on their behavior over time were then used to model their internal cross-strata. Both the bedform reconstructions from the cross-strata, and the models of the cross-strata from the modern dunes utilise the computer program of Rubin (1987). Although the Tsondab dune reconstructions are limited by both geographic extent and the quality of the outcrops, and the model for the internal structure of the modern dunes remains only a model until documented in the field, the treatment in this study does provide at least a preliminary basis for a comparison of Tsondab dunes to their modern counterparts.

TSONDAB SANDSTONE

Elim outcrop

The outcrop at Elim consists of a series of connected, near vertical walls (Walls A-H) of varying orientations, 2.5-7.5 m high, with a composite length of 76 m (Fig. 2). The outcrop is capped by an erosional surface littered with caliche, and overlain by submodern, vegetated, aeolian sands. The outcrop is clearly divisible into:

i) a lower series of sets bounded by prominent surface 1;

ii) a middle, completely bioturbated zone overlying the surface; and

iii) an upper single set bounded by surface 2.

Plant and animal bioturbation is prominent throughout the outcrop. In some cases, bioturbation is focused along foresets and surfaces, thereby enhancing these, but in other cases bioturbation has destroyed any primary structure. Avian eggshells found in this study and reported by Pickford *et al.* (1995) place the main body of the outcrop in the *Namornis oshanai* zone thought to be pre-Miocene, whereas the capping deflationary surface 2 yielded eggshells from *Struthio daberasensis* of a post-Miocene age.

The lower series of sets consists of sets of two different orientations. Sets with apparent dips to the left on Walls A-B on Fig. 2 consist of foresets dipping towards the south to south-southwest. Sets with apparent dips to the right on Walls B-H consist of foresets dipping towards the north-northeast to south-southeast. These sets of differing orientations overlap in Wall B to produce a 'zig-zag' arrangement. Where possible to determine with certainty, sets

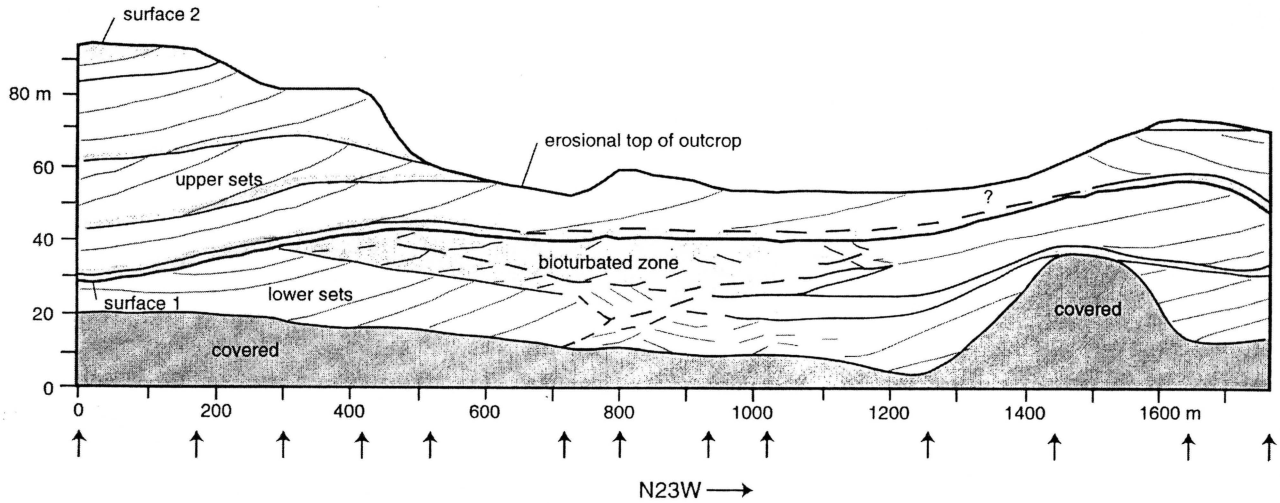

__Figure 3__. Cross-section of outcrop exposed in cliffs at Diep Rivier, indicated on Fig. 1 and discussed in the text. Arrows show the locations of the vertical sections. Foreset orientations are apparent for the section trend of N23W.

consist of grainflow strata with less abundant wind-ripple laminae. Scalloped bounding surfaces, dipping east to east-southeast, are evident within the lowest set on Walls C and E, and are probably present elsewhere but could not be identified with certainty.

Prominent surface 1 capping the lower series of sets is not horizontal, but rather is highest at the termination of Wall A, then progressively falls 3 m to its last clearly visible point on Wall E (note that in this, as in all cases for this study, it is assumed that there is no post-depositional tectonic dip). The surface is lost to bioturbation on Walls F-H. The bioturbated zone overlying this surface on Wall A is about 0.7 m thick, and the extent of bioturbation extending downward from this zone progressively increases to a maximum on Wall H. Bioturbation, therefore, is at a minimum at the highest point on bounding surface 1 and increases downslope.

The upper set of compound cross-strata, up to 1.6 m high, is poorly exposed and difficult to measure at Elim, but foresets generally dip to the east-northeast to east.

Diep Rivier outcrop

The outcrop at Diep Rivier has two parts. The main part of the outcrop is a cliff up to about 90 m high, and trending about N23W, of which 1760 m was traversed for this study (Fig. 3). The traverse was constructed by tracing beds and surfaces where possible on the steep cliff between 13 vertical sections indicated in Fig. 3. Beginning near the southern terminus of this outcrop and extending southward is a bench that exposes, in a horizontal section, several hundred metres of cross-strata (Fig. 4). These sets underlie those exposed in the cliff (Fig. 3), but they appear to be a continuation of sets exposed in the lower portion of the cliff. As at Elim, the Diep Rivier outcrop is clearly divided into:

i) a lower series of sets capped by prominent surface 1;

ii) a bioturbated zone overlying the surface; and

iii) an upper series of sets.

The caliche-littered erosional surface, surface 2, capping the upper sets occurs only in the southern portion of the cliff, with the outcrop progressively downcut to the north. Similarities to Elim in the stratigraphical zones are in agreement with the biostratigraphy, with the Diep Rivier outcrop yielding the pre-Miocene *Namornis oshanai* and *Struthio daberasensis*, occurring on deflationary surface 2 (Pickford *et al.*, 1995).

The lower series of sets along the Diep Rivier cliff consists of at least three large sets and a complicated array of small sets (Fig. 3). The lower large set, up to 20 m thick, exposed at the southern end of the traverse at Diep Rivier, shows cross-strata that exhibit a regular progression in foreset orientation, dipping toward the northeast (N55E) at the southern terminus of the traverse to culminate in dips towards the southeast (S52E) at the northern extent of the set. The two sets, each averaging about 20 m thick, in the northern portion of the traverse consist of a range of foresets dipping S65E to E. Both sets yield southward to tangential foresets marked by intense bioturbation at the lowest elevations of the sets. A bioturbated horizon about 1.5 m thick occurs between the two sets, and appears to thicken into a prominent 6 m thick bed in the centre of the outcrop. The central part of the outcrop is complex, made so by both intense bioturbation and a

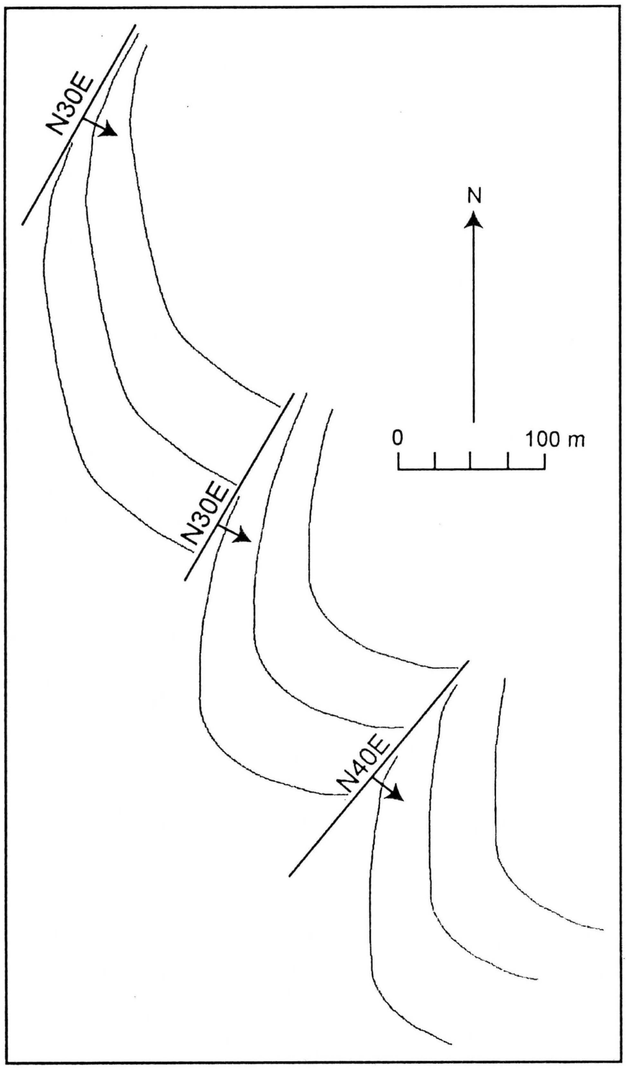

Figure 4. Horizontal section of cross-strata of the lower series of sets, as exposed on the bench at Diep Rivier.

complicated arrangement of sets. Although it was not possible to well-define the set arrangement, numerous foreset measurements in this zone showed a range (N25E-S80E) in part coincident with foresets measured in the lower set at the southern end of the traverse (N55E-S52E).

As at Elim, surface 1 bounding the lower sets is not horizontal, but rather irregularly rises about 20 m northward along the traverse (Fig. 3). The surface is overlain by about 1.5 m of intensely bioturbated sand, with the bioturbation extending down from the surface into the underlying sets.

Correlative to the upper set at Elim, at least four sets, ranging in thickness from 10 to 26 m, occur at Diep Rivier (Fig. 3). Foreset measurements from the sets at the southern portion of the traverse showed a range of S60-85E, but the set at the northern extreme of the traverse showed a dip toward the N25-46E. Stratification consists of mainly grainflow foresets, with lesser amounts of wind-ripple laminae, both of which toe tangentially to horizontal, heavily bioturbated bottomsets up to 1.5 m thick. Bounding surfaces occur within the sets, and measurements showed that their orientations are similar to the foresets.

Figure 4 shows the three best sets exposed in horizontal section along the bench at Diep Rivier. Sets are about 200 m wide, and bounded by surfaces trending N30-40E and dipping to the southeast. The foresets define nearly full, but yet asymmetric crescent shapes, with the foresets tangential to the bounding surface to the north, but truncated to the south by the subsequent surface. The mean foreset dip direction is about N55E, with the range of dip directions of N to S70E defining the set curvature.

Bioturbation at both the Elim and Diep Rivier outcrops can be summarised as:

i) intense zones overlying the prominent bounding surfaces;

ii) extending from the surfaces into the underlying sets;

iii) intense zones that interbed with foresets; and

iv) pervasive throughout the cross-strata.

No effort was made here to specifically identify the origins of the bioturbation, except to observe that it is very varied in size and structure, with clear examples of calcrete-cemented roots and termite nests (Fig. 5). In a more comprehensive study of the trace fossils, Ward (1988) found numerous similarities between those within the Tsondab Sandstone and traces made by organisms inhabiting the modern sand sea.

Overall aeolian system interpretation

The distance separating outcrops at Elim and Diep Rivier, as well as the relatively broad range of the biostratigraphic markers, limit the certainty of any stratigraphical correlations. However, these outcrops are tentatively correlated because:

i) the style of cross-strata in the lower and upper series of sets at both outcrops is very similar;

ii) both series are bounded by prominent surfaces with their associated bioturbation zone or paleosol horizons; and

iii) both outcrops yield the same avian eggshells.

Given (1) the prominence of surfaces 1 and 2; (2) the palæotopographic relief of the surface capping the lower sets; (3) the overlying bioturbated or palæosol zones; and (4) the somewhat differing styles of cross-strata in the lower versus the upper series of sets, in this study the surfaces are interpreted as sequence boundaries ('super surfaces' of Kocurek and Havholm, 1993). Because the origin of surfaces 1 and 2 is probably climatic (i.e. more humid conditions), these surfaces are likely of sand

Figure 5. Extensive root structures and bioturbation, which characterises much of the Tsondab Sandstone.

sea extent. Correlation of the aeolian units bounded by these surfaces does not necessarily mean that these belong to synchronous dunes, but rather only that they represent similar dune development within the same aeolian sequence.

Put into a genetic context, a period of accumulation of the lower series of sets occurred and was then terminated with the development of surface 1. This surface must represent a period of dune stabilisation with enhanced vegetation. Development of bioturbated accumulations above the sequence surface shows that some transport of sand occurred during this period of plant and animal colonisation, but it is not clear if this accumulation represents aeolian transport or a redistribution of sands by subaqueous and other processes with overall modification of the surface. Sections at both Elim and Diep Rivier show that bioturbation was influenced by the palæorelief, with a general trend towards increased bioturbation in topographic low areas, especially at Elim. The palæorelief itself may represent dune palæotopography, but with a significant degree of surface modification. It can be argued at Diep Rivier that the surface high (north end of traverse in Fig. 3) occurs over dune relief, then loosely parallels dune topography downward to where foresets intertongue with bioturbated bottomset deposits. At Elim, however, the surface does not convincingly show dune palæorelief.

Aeolian accumulation commenced again with the upper series of sets on top of the irregular relief of surface 1. This period of accumulation ended with development of surface 2, again believed to represent a sequence boundary, in this case particularly characterised by caliche development. A combination of greater accumulation within topographic low areas on surface 1, and varying depths of erosion in the formation of surface 2, probably account for the preservation of a single set in the upper series at Elim and multiple sets at Diep Rivier. The surface does slope in general to the north at Elim, whereas at Diep Rivier the surface is present only in the southern extreme of the traverse, with subsequent erosion truncating the section to the north.

The cross-strata in the Tsondab Sandstone are remarkable in the extent of plant and animal bioturbation preserved within the cross-strata. Although the sequence boundaries show enhanced bioturbation, it is obvious in outcrop that flora and fauna were prevalent throughout accumulation of the cross-strata, and that the Tsondab bedforms were vegetated to some significant degree during their construction and accumulation. Although the extent to which vegetation controlled accumulation or dune morphology cannot be determined with any certainty, the Tsondab aeolian system is arguably an ancient example of a stabilising aeolian system. Stabilising aeolian systems are defined by Kocurek and Havholm (1993) as those in which dune form and accretion over time is largely a function of stabilising agents such as vegetation.

Bedform reconstruction

Lower series of sets

Reconstruction of the bedforms that gave rise to the cross-strata in the Tsondab Sandstone is greatly hindered by both the paucity of large outcrops and the relatively poor preservation of the dune primary structures. Given the correlations between Elim and Diep Rivier, however, a composite template can be assembled that, if not specific to a particular dune morphology and behaviour, does at least place constraints on possible interpretations, especially as regards how similar these Tsondab bedforms were to modern Namib dunes.

For the lower series of sets, the most diagnostic outcrop is the horizontal section at Diep Rivier (Fig. 4), which shows features best explained as formed by the presence of a main bedform with superimposed, migrating scour pits or bedforms. For the case of superimposed dunes (e.g. Rubin 1987, fig. 46), the migration of the main bedform and the superimposed features creates scour pits that have both crest-perpendicular and crest-parallel components of migration with respect to the main bedform. In a horizontal section, the scour pits are truncated in the direction of migration of the main bedform as the full crescent shape or trough of the scour pits is truncated by the passage of the next upwind scour pit, which migrates parallel to the previous scour pit but is displaced laterally by migration of the main bedform. The degree of displacement, reflected in both the extent of trough truncation and the trace of the bounding surfaces in the horizontal section, is determined by the relative migration rates of the main bedform and the superimposed features. A similar structure can be generated by the migration of sinuosity parallel to the crest of the main bedform, with or without net lateral migration of the main bedform (e.g. Rubin 1987, fig. 55).

Assuming originally symmetrical crescent shapes before truncation, the crescent sets in horizontal section were in the order of 300 m wide. The average dip direction of the northeast trending bounding surfaces is S55E, whereas the average foreset dip direction is about N55E. As shown by Rubin and Hunter (1983), if the structure is created by the migration of superimposed dunes upon a larger bedform, then the line of intersection on a stereonet of the plane of the bounding surface and that of the cross-strata defines the trend of the superimposed dunes, here yielding a trend that is approximately east, with migration to the north. From this outcrop, however, it is not possible to determine the morphology and orientation of the main bedform, but in order to produce the structures present, there must have been net migration of the main bedform towards a general easterly direction. Assuming that migration is perpendicular to the crest, then the main bedform trended roughly north-south. If the structure is produced by the migration of sinuosity along the axis of a dune, the line of intersection bisects the sinuosity or is perpendicular to the overall dune trend, in this case yielding a dune axis trending roughly north-south. In either case, the predicted cross-sectional views from the horizontal section are scalloped sets in a general east-west section, and sets climbing to the north in a general north-south section. The precise shape of the sets would be determined by the relative rates of migration of the main and superimposed features and their orientations. A component of eastward migration for the main bedform does not, however, preclude a component of crest-parallel migration or extension as well.

Although far larger than the horizontal section at Diep Rivier, the cross-section revealed in the traverse (Fig. 3) is less forthcoming in interpretative value. For the total traverse of the lower series of sets, foreset dip directions measured range from N25E to S52E. It is reasonable to believe that this range reflects the foreset curvature evident in the horizontal section, which showed a measured range of N to S70E. Changing foreset dip directions along the traverse indicate that the traverse trend is not parallel to the migration direction of the scour pits. Although the complex central portion of the traverse bears a similarity to the zig-zag pattern at Elim, foresets at Diep Rivier do not show the south to south-southwest dip directions at Elim, but rather belong to the family of sets characterised by the large sets at Diep Rivier.

The outcrop at Elim potentially complicates the emerging reconstructed bedform for the lower series of sets because of the presence of the sets with foresets dipping to the south to south-southwest. The predominate sets at Elim, with foresets dipping to the north-northeast to south-southeast belong to the same class of sets as those at Diep Rivier. Unfortunately, the sort of revealing horizontal section present at Diep Rivier does not occur at Elim. The 'zig-zag' structure at Elim, however, can reasonably be taken as having been produced by the shifting point of division over time of a central point, either the crest or trough of the main bedform, with the alternate orientations of cross-strata on its two flanks. This type of structure could be produced by the along-crest migration of sinuosities superimposed upon a larger bedform (e.g. Rubin 1987, fig. 55), or by superimposed dunes of different orientations and migration directions on alternate flanks of the main bedform. The predominance of sets with dips toward the northeast, however, shows that net migration of the main bedform in a general easterly direction occurred, thereby causing erosion of the western flank sets (e.g. see example of reversing bedform with net migration in Rubin 1987, fig. 21).

Upper series of sets

The upper series of sets at Elim and Diep Rivier are puzzling because, whereas they show a range of foreset dip directions (S60-85E being well represented, and N25-46E in one set) that completely fall within the total range of that of the lower series of sets (N to S52E), they exhibit few of the complexities evident in the lower series. Internal set bounding surfaces, because they are concordant to the foresets, are best interpreted as reactivation surfaces on the lee face, and not the product of the migration of superimposed bedforms (Rubin and Hunter, 1983). Viewed in isolation, these sets could be interpreted as representing relatively simple crescent bedforms with a curvature defined by the range of foreset dip directions and migrating east-northeast. Alternately, the bedforms could be the same as the more complex bedforms represented by the lower series, but in which the complexities of the bedform evident in the lower series of sets have been lost. This loss of bedform history would occur with eastward migration of the main bedform that results in erosion of all sets except for those in the net migration direction.

MODERN LINEAR DUNE MODELING

Input variables

The predominant dune type of the present-day Namib Sand Sea is very complicated, but can generally be characterised as a sinuous-crested linear dune with varying development of smaller, superimposed crescent dunes and, less commonly, star dunes (i.e. complex linear dunes, see Lancaster, 1989, fig. 7; Rubin, 1987, fig. 76). In order to model the internal structure of these dunes using the computer program of Rubin (1987), it is necessary to input values into the program that describe both the morphology of the dunes and their behaviour over time.

Linear dune height is taken here as 100 m based upon a range of 80-120 m with a mean of 99 m (Lancaster, 1989, p. 35). Dune trend, as measured from satellite images (e.g. Lancaster, 1989, photo 1), varies somewhat through the sand sea with dune curvature, but is in general north-south, with the average taken here as N5W. Dune asymmetry is 0.18 (where zero is perfectly symmetric and 1 is strongly asymmetric) based upon topographic maps and surveyed profiles (Lancaster, 1989, fig. 11; see also Livingstone, 1986, 1989), with the overall steeper face towards the west. The upper crestal zone of the dunes, however, is steeper on the eastern side, but then proceeds downward to an extensive, lower-angle dune flank. Seasonal, reversing migration of the crestal portion of the dunes also occurs (Livingstone, 1989). Dune spacing varies from 1600-2800 m with a mean of about 2100 m (Lancaster, 1989, p. 35), yielding an average spacing to height ratio of 21:1. The average width of the dunes is 886 m (Lancaster, 1989, p. 35), or about 40% of the dune spacing is occupied by dunes. From aerial photos, the spacing of the sinuosity of the crestline varies from less than 200 m to over 450 m, with 300 m used here. The amplitude of this sinuosity is in the order of 100 m. The range of development of superimposed crescenc and star dunes is from a complex of two sets orientated oblique to the main linear dunes (see Rubin, 1987, fig. 76) to essentially an absence of superimposed dunes. In general, the superimposed crescent dunes are best developed on the more-gently sloping eastern plinths of the main linear dunes, but they are small in size (2-10 m, Lancaster, 1989, p. 31) in comparison to both the main bedform and its sinuosity. Because of: (1) the relative small size of the crescentic dunes; (2) the lack of consistent development of both crescent and star elements; (3) an uncertainty of whether the presence of the superimposed dunes is inherent to the overall linear dunes through time or only a more recent modification; and (4) modelling only the most basic, large-scale internal structure in this study, the superimposed dunes have been omitted from this model.

The annual potential sand transport rose for the Namib varies over the sand sea (Lancaster, 1989, fig. 47), but basically is the product of south-southwest winds (taken here as from S20W) that predominate from September-April, and northeast winds (taken here as from N60E) that become more important during the winter with a maximum in June or July. For most of the sand sea, and increasing toward the coast, the south-southwest winds predominate over the northeast winds. The dunes are oblique to both average wind directions: by 65° for the northeast winds, and 25° for the south-southwest winds. Because of the oblique incidence angle of the primary winds with respect to the dune trend, secondary flow on the alternate lee faces should be deflected alongslope (e.g. Tsoar, 1983; Havholm and Kocurek, 1986; Kocurek, 1991), yielding: (1) northward winds along the eastern flank for south-southwest winds; and (2) southward winds along the western flank for northeast winds.

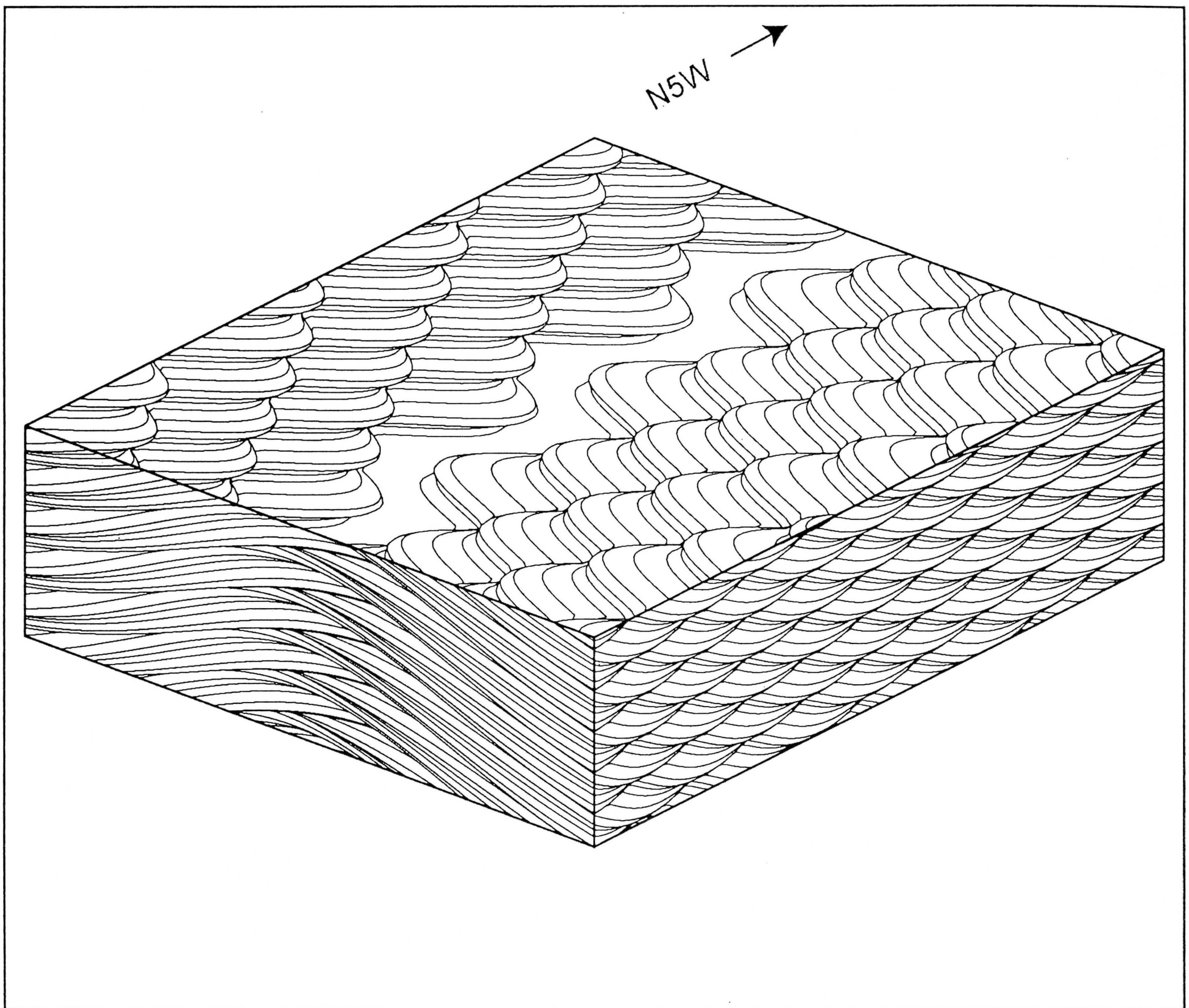

Figure 6. *Modern linear dune Model I in which the dunes grow vertically and laterally, and extend northward with the migration of sinuosity along the crest of the bedform.*

The most basic question as to the dune behaviour over time is whether or not the linear dunes have a component of net lateral migration. Northward extension of the dunes (i.e. parallel to their crestlines) towards the Kuiseb River has been well-documented (see data and review in Ward and von Brunn, 1985), but lateral migration is less certain. In one unpublished ten-year study (1979-1989) of a dune located about 15 km from the Gobabeb Research Unit near the Kuiseb River (J. Ward, *pers. comm.* 1999), the base of the western slope of the dune migrated eastward about 1 m, while 4-6 m of northward extension occurred. In another study (1980-1992) in the same area, no net movement of the dune base was seen (Livingstone, 1993). In order to contrast internal dune structure, two models have been generated–one without net lateral migration and one with net lateral migration. Although both models are complex, in reality the actual structure of the dunes is almost certainly even more complicated because of the presence of superimposed dunes. In addition, it is probable that the present-day dunes of the Namib are not wholly the product of the modern wind regime, but rather a composite of modern and more ancient regimes.

Model I envisions the simpler case in which there is no net lateral migration of the main linear dunes. In this view, the deflected alongslope secondary flow causes nearly all sand movement to be parallel to the axis of the dune. Crestal sinuosity, however, migrates along the axis. Because of the predominance of the south-southwest winds, migration should be to the north, but with seasonal reversals caused by the northeast winds. For the model, a net northward migration of 1 m a^{-1} was assumed, with components of 1.5 m to the north and 0.5 m to the

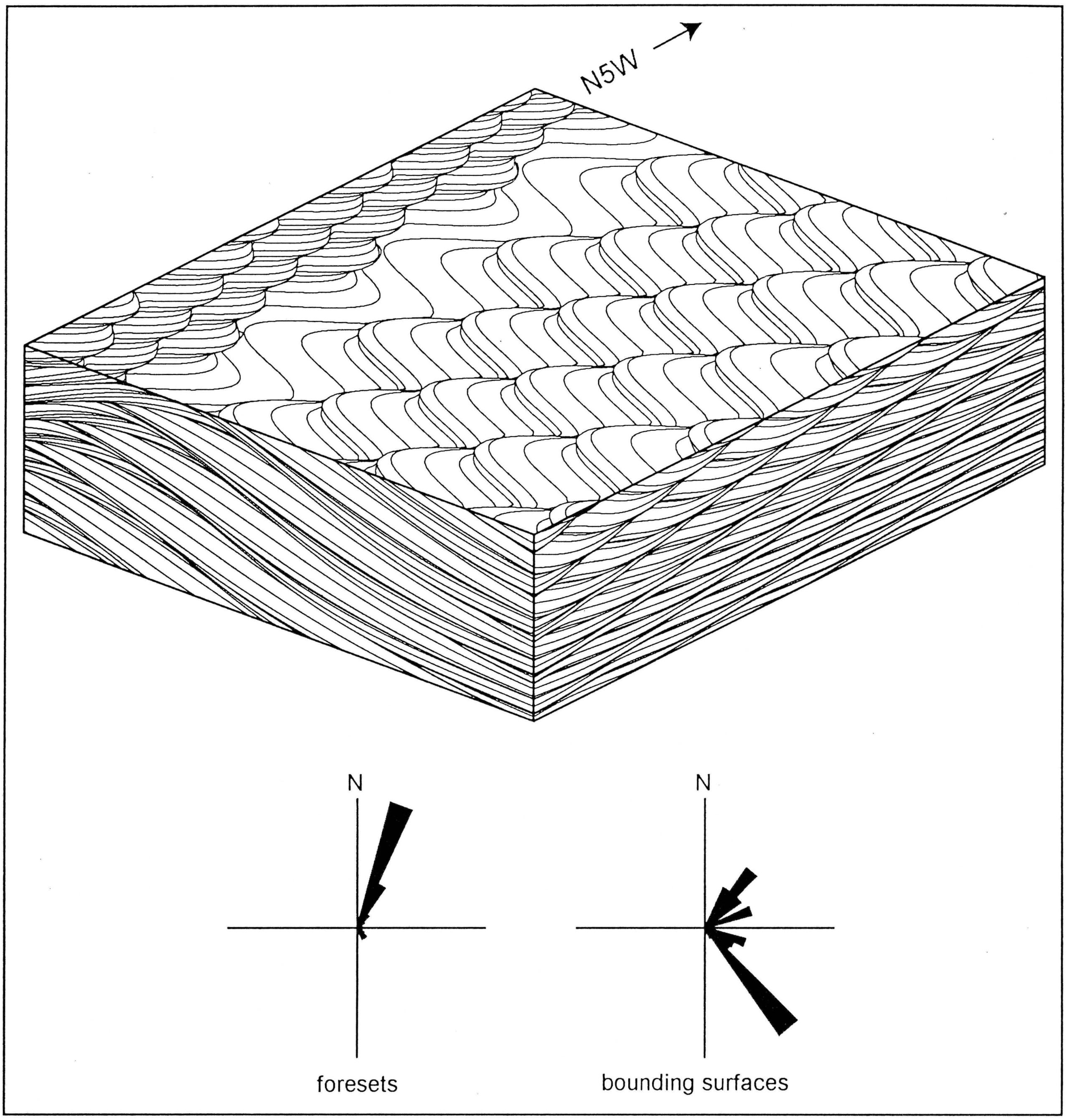

Figure 7. Modern linear dune Model II in which the dunes grow vertically and extend northward with the migration of sinuosity along the crest of the bedform, but also with net lateral bedform migration to the east. Orientations of foresets and bounding surfaces are for the eastern flank of the bedform.

south during one wind yearly cycle. In this model, over time, the dune grows vertically and laterally on both flanks through the passage of sinuosity along the crest, while extending in an alongcrest direction to the north.

Model II also assumes a northward extension of the dune, with sinuosity migration identical to Model I, but with a net lateral migration for the main bedform to the east. In Model II, therefore, there is one component of dune behaviour in which the dunes extend northward alongcrest with a corresponding component of sinuosity migration in response to the strongly oblique winds from the south-southwest, albeit with reversals as in Model I. A second component of migration is a net eastward displacement of the dunes caused by the predominant south-southwest winds, although there is also reversals in lateral migration which occur during the wind seasons. In this view, the predominance of the south-southwest winds causes more sand to be transported to the eastern flanks of the dunes than is transported to the western flanks of the dunes by the northeast winds, thereby causing a net eastward migration. The predominance of the south-southwest winds is taken as over-riding the fact that the more transverse incidence angle of the northeast winds might argue for a westward net lateral migration. Selection of migration rates is arbitrary but relative: the rates of sinuosity migration in Model I were assumed, with the assumption of a smaller net lateral migration rate of 0.3 m a^{-1} for the dune overall, with components of 0.4 m a^{-1} to the east and 0.1 m a^{-1} to the west. Whereas the assumed rates are reasonable for dune migration, it should be noted that these rates will affect the details of the structures but not the basic configuration of the sets.

Model results

Simulation of Model I (sinuous linear dunes with no net lateral migration) is shown in Fig. 6. The basic structure results from the lateral migration of the sinuosity, while the overall bedform grows vertically and expands laterally. The structure has bilateral symmetry, caused by the passage of the sinuosity, which, in respect to an imaginary line down the axis of the dune, occurs on alternate sides of this line. This imaginary line bisects the zig-zag structure that characterises cross-strata in the cross-section perpendicular to the dune axis. In the cross-section parallel to the dune axis, climbing sets occur, each of which marks the migration of the concave portion of the sinuosity. In horizontal section, these sets are troughs truncated in the direction of dune lateral growth by the passage of the erosional convex portion of the sinuosity. The distribution of forests reflects the bilateral symmetry of the sets, with equally great clusters of steep foresets dipping to the east and west. Reactivation surfaces or bounding surfaces within the sets are prominent and caused by the coincidence of seasonal reversals in the migration direction of the sinuosity and the bedform overall.

Simulation of Model II (sinuous linear dunes with net lateral migration to the east) essentially results in net accumulation of the eastern half of the structure shown in Model I, with the important difference that sinuosity, while migrating alongcrest towards N5W, is displaced on the eastern flank of the bedform by the migration of the bedform towards N85E (Fig. 7). With net lateral migration, the western flank of the structure is ultimately removed as it acts as the erosional stoss slope of the overall bedform over time. Remnants of the western flank remain only where not completely removed by deflation, producing a strongly unidirectional structure with the mean foreset dip direction towards the north-northeast, whereas the set bounding surfaces have a mean dip towards the east.

COMPARISON OF TSONDAB AND MODERN DUNE STRUCTURES

Although there are important similarities between the Tsondab set architecture and that of the models of the modern linear dunes, it has to be concluded that the Tsondab dunes were not simply ancient counterparts of the modern dunes, at least as depicted by the models in this study.

Points of similarity include:

i) a northward trend of both Tsondab and modern dunes;

ii) a general eastward migration direction for the Tsondab dunes and those in Model II;

iii) the development of scalloped cross-strata at Diep Rivier and in both models;

iv) a reconstructed scour trough width of about 300 m in the Tsondab sets that compares well to the spacing of sinuosity along the crests of the modern dunes; and

v) development of reactivation surfaces within sets of the Tsondab and model dunes because of seasonal changes in wind direction.

These similarities argue that some key elements of the basic configuration of the sand sea during Tsondab times remain intact in the modern sand sea.

Differences between the models in this study for the modern linear dunes and Tsondab outcrops reside in foreset and bounding surface orientations, as well as the gross appearance of the structures. There is little from Model I, in terms of the bilateral symmetry of the overall structure or the foreset distribution, that compares with the Tsondab sets. The Tsondab dunes were not stationary, but rather had a strong component of general eastward migration, as with Model II. However, the climbing sets shown in the right-hand cross-section of Model II do not compare well with the Diep Rivier elongated troughs, even when the model is rotated 18° so that the cross-section is parallel to the trend of the outcrop at Diep Rivier. Tsondab foresets dip in a range of north to S70E with the trough axis trending at N55E, whereas Model II foresets dip in a range of N15E to S30E with the trough axis trending at N25E. Scalloped set bounding surfaces in the Tsondab strike at N35E, whereas those in Model II strike at N10E. Tsondab cross-strata at Elim record minority sets with foresets dipping toward the south-southwest, which have no counterparts in Model II;

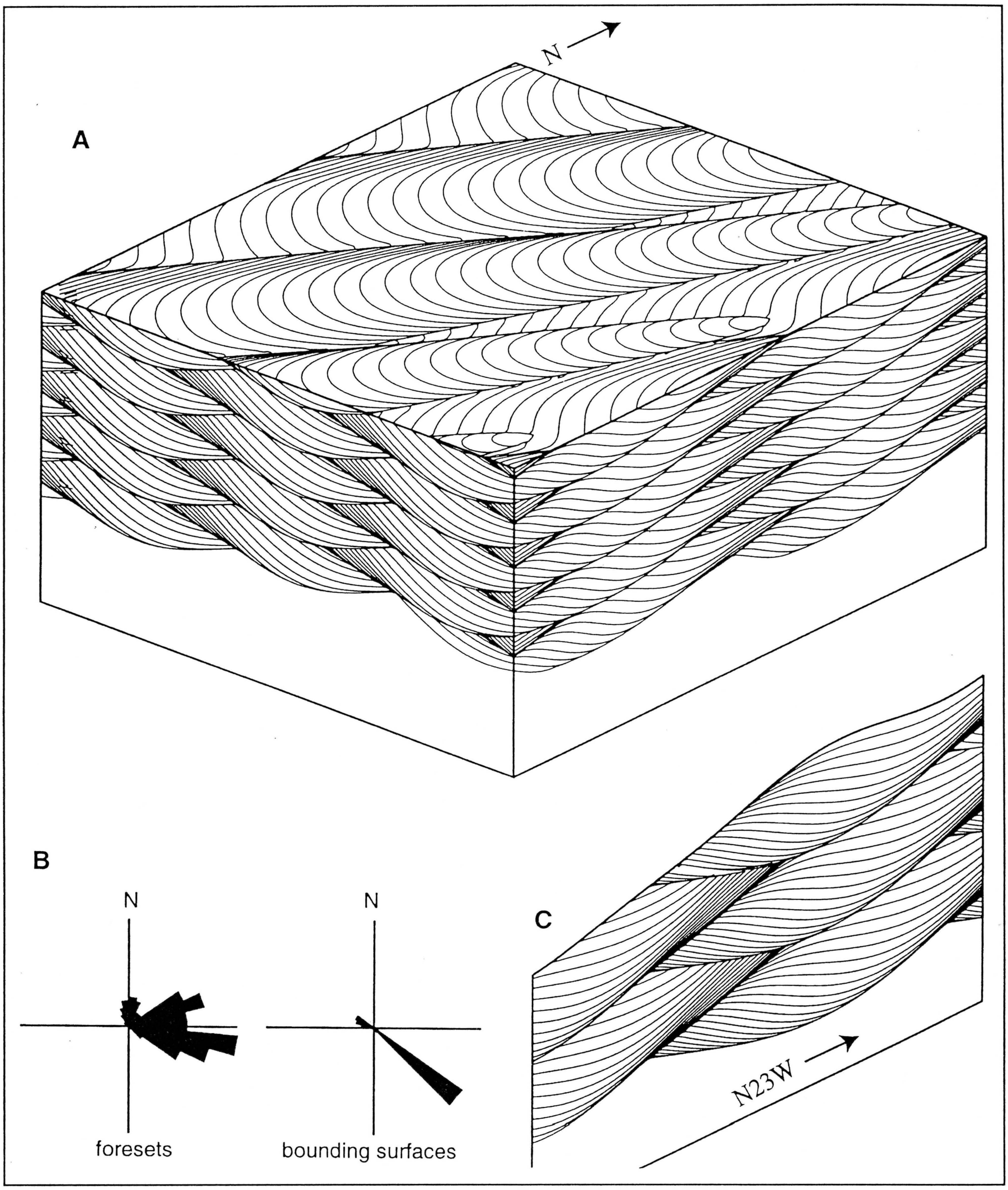

Figure 8. Model that best satisfies the Tsondab outcrop data in which northward-migrating, superimposed, large dunes occur on the eastern flank of a north-south trending main bedform. (A) Model in block diagram. (B) Orientation of foresets and bounding surfaces. (C) Model wall rotated to the same orientation as the cliff at Diep Rivier (Fig. 3).

In order to test whether the Model II structure could conform to the Tsondab structure with reasonable changes in orientation and rates of migration, numerous simulations were made, but these failed to achieve agreement in either structure or orientation.

In addition to a mismatch in set architecture, a fundamental difference between the Tsondab Sandstone and the modern sand sea is the degree of vegetation that must have characterised the Tsondab sand sea. Currently, vegetation characterises some

Namib dunes, especially the linear dunes in the eastern part of the sand sea (see photo 6 in Lancaster, 1989), and very significant increases in vegetative cover have been documented within the Namib Desert during wetter years (Seely and Louw, 1980). However, the extent of bioturbation evident in the Tsondab is simply not manifested within the modern dunes, suggesting that the Tsondab dunes did not form in the hyperarid climate that characterises the Namib today.

TOWARDS A TSONDAB MODEL

Although the current level of outcrop data is insufficient to generate a unique Tsondab bedform model, and it is possible that the nature of the Tsondab outcrops preclude this possibility, a tentative model can be advanced that does largely satisfy the outcrop data collected. Because of the failure to reproduce the Tsondab structure by the migration of alongcrest sinuosity, the alternative model suggested by the scallops at Diep Rivier was tested, namely a main bedform trending north and migrating to the east, with superimposed dunes trending east-west and migrating to the north along the eastern flank of the main bedform. In order to achieve the large, nearly complete troughs evident in the Diep Rivier horizontal section and traverse, the Tsondab troughs must have been relatively large in comparison to the main bedform and not migrating to the north appreciably faster than the main bedform was migrating to the east (e.g. Rubin, 1987, fig. 46j).

The most successful simulation used a main-to-superimposed dune migration speed ratio of 1:1.5 (Fig. 8A). This simulation yields a range of foresets of N to S50E that defines crescent shapes with trough axes trending to the northeast, and set bounding surfaces striking at N40E (Fig. 8B). These orientations compare well to the outcrop data. In horizontal section, crescent shapes are nearly fully yet asymmetric, with foresets tangential to the bounding surfaces to the north but truncated to the south, as with the horizontal section at Diep Rivier. When the structure is rotated to have the right-hand cross-section parallel to the section at Diep Rivier (Fig. 8C), the elongate troughs and zig-zag features (caused by the intersection of troughs of the same orientation) apparent at Diep Rivier are seen in the model. For simplicity, reactivation surfaces have been omitted from the model. In addition, no attempt was made to model the south to south-southwest dipping foresets at Elim, but these may indicate that superimposed dunes also occurred on the western flanks of the main bedforms and migrated to the south in response to alongslope directed winds from the northeast. These sets were only rarely preserved because of the strong eastward migration of the overall dune mass.

Although the simulation of the Tsondab dunes that best approximates the outcrop structure uses a fundamentally different model than that used in the simulations of the modern dunes, morphologically these are very similar bedforms. The Model II linear dunes trend north-south and migrate to the east, with the primary structure created by the alongcrest migration of sinuosity to the north, whereas the role of superimposed dunes *per se* was neglected because they seem minor in comparison to the scale of the sinuosity. The best Tsondab model dunes similarly trend north-south and migrate to the east, but sinuosity is neglected with emphasis rather upon the migration of superimposed dunes that migrate alongslope to the north. Nothing in this Tsondab model precludes linear dunes. On the one hand, differences in the models could reflect genuine bedform differences, whereby the Tsondab dunes were straight-crested with large, prominent, superimposed dunes in comparison to the modern linear dunes. On the other hand, these differences could simply reflect how these complicated modern dunes were characterised.

CONCLUSIONS

The Tertiary Tsondab Sandstone Formation is important to understanding the evolution of the Namib Desert because it is one of the few stratigraphical units in the region that directly demonstrates that aeolian conditions occurred as early as the Miocene or before in the area now occupied by the modern sand sea. Based upon the outcrops examined, Tsondab conditions were more humid than at present, with active dunes vegetated to some significant degree. There are at least two periods, represented by likely formation-wide sequence bounding surfaces, in which sand sea accumulation ceased and the surface was stabilised by vegetation and reworked to some extent. A preliminary model based upon outcrop data suggests that the Tsondab dunes were north trending features migrating to the east, with relatively large and slow-moving dunes superimposed on the eastern flanks of the main bedforms and migrating to the north. Limited data suggest that superimposed dunes also occurred on the western flanks of the main bedforms and migrated to the south, but that these sets were largely removed by erosion associated with the net eastward migration of the overall bedform. The trend of the Tsondab dunes, which is the same as the modern dunes, and their at least gross morphological similarity to the modern linear dunes suggest that similar wind conditions existed during Tsondab times.

ACKNOWLEDGEMENTS

This research was funded in part by a grant from the National Geographic Society (NGS 4821-92). The authors thank the Department of Environment and Tourism, Government of the Republic of Namibia, for permission to conduct research in the Namib-Naukluft Park. Martin Pickford and Roger Swart were extremely helpful in their assistance with the Tsondab outcrops. The owners of the Diep Rivier farm allowed the authors to conduct research on their property. Reviews of this paper by Ron Blakey and John Ward were very constructive. The authors especially thank John Ward for sharing his knowledge and data regarding aspects of the Tsondab Sandstone Formation and the modern Namib Desert.

Editorial handling - S. Goodman

REFERENCES

Barnard, W.S., 1973. Duinformasie in die Sentrale Namib. Tegnikon 22, 2–13.

Besler, H., Marker, M.E., 1979. Namib Sandstone: a distinct lithological unit. Transvaal Geological Society South Africa 82, 155–160.

Havholm, K., Kocurek, G., 1986. A preliminary study of the dynamics of a modern draa, Algodones, southeastern California, USA. Sedimentology 35, 649–669.

Kocurek, G., 1991. Interpretation of ancient eolian sand dunes. Annual Review Earth Planetary Science 19, 43–75.

Kocurek, G., Havholm, K., 1993. Eolian sequence stratigraphy - a conceptual framework. In: Weimer, P., Posamentier, H.W. (Eds.), Siliciclastic Sequence Stratigraphy. American Association Petroleum Geologists Memoir 58, pp. 393–409.

Lancaster, N., 1989. The Namib Sand Sea: Dune Forms, Processes and Sediments. A.A. Balkema, Rotterdam, 180p.

Livingstone, I., 1986. Geomorphological significance of wind flow patterns over a Namib linear dune. In: Nickling, W.G. (Ed.), Aeolian Geomorphology. Binghamton Symposia in Geomorphology 17, pp. 97–112.

Livingstone, I., 1989. Monitoring surface change on a Namib linear dune. Earth Surface Processes Landforms 14, 317–332.

Livingstone, I., 1993. A decade of surface change on a Namib linear dune. Earth Surface Processes Landforms 18, 661–664.

Martin, H., 1950. Sudwestafrika. Geologische Rundschau 83, 6–14.

McKee, E.D., 1982. Sedimentary structures in dunes of the Namib Desert, South West Africa. Geological Society America Special Paper 188, 64p.

Ollier, C.D., 1977. Outline geological and geomorphological history of the Central Namib Desert. Madoqua 10, 207–212.

Pickford, M., Senut, B., Dauphin, Y., 1995. Biostratigraphy of the Tsondab Sandstone (Namibia) based on gigantic avian eggshells. Geobios 28, 85–98.

Rubin, D., 1987. Cross-bedding, Bedforms, and Paleocurrents. SEPM Concepts in Sedimentology and Paleontology 1, 187p.

Rubin, D., Hunter, R., 1983. Reconstructing bedform assemblages from compound crossbedding. In: Brookfield, M.E., Ahlbrandt, T.S. (Eds.), Eolian Sediments and Processes. Elsevier, Amsterdam, pp. 407–427.

Seely, M.K., Louw, G.N., 1980. First approximation of the effects of rainfall on the ecology and energetics of a Namib Desert dune ecosystem. Journal Arid Environments 3, 25–54.

Tsoar, H., 1983. Dynamic processes acting on a longitudinal (seif) sand dune. Sedimentology 30, 567–578.

Ward, J.D., 1987. The Cenozoic succession in the Kuiseb Valley, central Namib Desert. South West Africa Geological Survey Memoir 9, 124p.

Ward, J.D., 1988. Eolian, fluvial and pan (playa) facies of the Tertiary Tsondab Sandstone Formation in the central Namib Desert, Namibia. Sedimentary Geology 55, 143–162.

Ward, J.D., Corbett, I., 1990. Towards an age for the Namib. In: Seely, M.K. (Ed.), Namib Ecology: 25 Years of Namib Research. Transvaal Museum Monograph 7, pp. 17–26.

Ward, J.D., Seely, M.K., Lancaster, N., 1983. On the antiquity of the Namib. South African Journal Sciences 79, 175–183.

Ward, J.D., von Brunn, V., 1985. Sand dynamics along the lower Kuiseb River. In: Huntley, B.J. (Ed.), The Kuiseb Environment: The Development of a Monitoring Baseline. South African National Science Programme Report 106, pp. 51–72.