INTERNATIONAL JOURNAL OF CLIMATOLOGY, VOL. 10, 63–77 (1990) 551.553.1(688)

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THERMO-TOPOGRAPHICALLY INDUCED BOUNDARY LAYER OSCILLATIONS OVER THE CENTRAL NAMIB, SOUTHERN AFRICA

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Received 11 October 1988
Revised 31 January 1989

ABSTRACT

Diurnal and seasonal oscillations of the atmospheric boundary layer over the central Namib Desert on the west coast of southern Africa are examined. Both the vertical and horizontal structure of the thermo-topographic airflow are contrasted with similar wind systems occurring over Natal on the east coast of the subcontinent. Thermo-topographic airflows over the central Namib are found to have a regional significance frequently equalling or exceeding that of the general circulation. The strength, depth, and unusually clearly defined diurnal and seasonal oscillations of these winds render the central Namib a unique area for the study of boundary layer oscillations.

KEY WORDS Boundary layer oscillations Thermo-topographic airflows Sea-breezes Mountain–plain winds Plain–mountain winds Central Namib Desert Southern Africa

INTRODUCTION

Local climates of coastal and mountain areas are characterized by the modification of synoptic-scale airflow by thermally and topographically induced diurnal oscillations of the lower boundary layer. Numerous observational studies of the sea/land-breezes, local mountain/valley winds and regional mountain/plain winds which constitute these oscillations have led to an improved theoretical understanding of individual thermo-topographic wind systems (Defant, 1958; Flohn, 1969; Atkinson, 1981; Yoshino, 1981; Sturman, 1987). Numerical modelling of these systems has further contributed to the delineation of their characteristics and dynamics (see the reviews of Atkinson (1981, 1983) and Pielke (1984); and also Huss and Fetiles (1981), Kikushi *et al.* (1981), Garrett (1983), Rotunno (1983), Lee and Kau (1984), McNider and Pielke (1984), Mass and Dempsey (1985), Abbs (1986), Kitada *et al.* (1986), and Egger (1987)). The complex wind regimes produced by the interactions of topographic and thermal effects on local and regional scales with each other and with the synoptic circulation, however, are not as well documented.

Empirical studies in which interactions among various thermo-topographic boundary layer wind systems and larger-scale airflows are described have been undertaken for areas such as Indonesia (van Bemmelen, 1922), Austria (Defant, 1949), California (Edinger and Kao, 1959; Frenzel, 1962), Vermont (Davidson and Rao, 1963), East Africa (Flohn, 1965), Washington (Buettner and Thyer, 1966), Arabia (Steedman and Ashour, 1976), Venice (Camuffo *et al.*, 1979; Camuffo, 1982), Israel (Skibin and Hod, 1979; Bitan, 1981; Goldreich *et al.*, 1986), Australia (Kamst *et al.*, 1980), and the Canterbury Plains of New Zealand (Sturman and Tyson, 1981; McKendry, 1983; McKendry *et al.*, 1986). Such studies are largely based on surface wind information, augmented by limited vertical observations. In southern Africa the nature and interactions of local and regional thermo-topographic winds over the eastern plateau slopes of Natal have been well considered in this way (Tyson, 1966, 1967, 1968a, b; Preston-Whyte, 1969, 1974), and a simple model has been postulated to account for the diurnal oscillation of winds in the lower boundary layer between the Drakensberg Escarpment and the east coast (Tyson and Preston-Whyte, 1972; Preston-Whyte and Tyson, 1988). Local wind regimes over the central Namib Desert on the southern African west coast are not as well described and

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understood, although Goldreich and Tyson (1988) have shown that diurnal boundary layer processes dominate the near-surface windfield there in all seasons. Characteristics of the sea-breeze have been documented for this coast (Jackson, 1942, 1954); further inland, topographically and thermally induced air movements have been identified only from surface data (Tyson and Seely, 1980).

In this paper, diurnal and seasonal oscillations of the boundary layer over the desert of the central Namib will be examined. Both the vertical and horizontal structures of the thermo-topographic airflow will be considered and contrasted with similar wind systems occurring at approximately the same latitude over Natal on the east coast of southern Africa. Interactions among the various meso-scale and synoptic wind systems will also be discussed.

PHYSICAL SETTING AND GENERAL AIRFLOW CLIMATOLOGY

The sparsely vegetated central Namib Desert is bounded by the cold Atlantic Ocean to the west and a dissected western plateau slope beneath the Escarpment (which has an average altitude of 1500 m, rising to 2300 m in places) some 160–180 km to the east (Figure 1a). The deeply incised Kuiseb River valley forms a boundary between the relatively flat gravel plains to the north and the coast-parallel linear dunes of the sand sea to the south. A research station is situated in the Kuiseb Valley at Gobabeb (Figure 1a), which is 56 km inland from the coast and is where the majority of the measurements included in this paper were made. The north-west–south-east trend of the Kuiseb Valley, from near the coast to some distance beyond Gobabeb, changes to north-east–south-west nearer to the Escarpment (Figure 1a).

Physical contrasts between the western and eastern plateau slopes of southern Africa are great (Figure 1b). The warm waters of the Agulhas Current and well-vegetated eastern coastal margins of Natal contrast markedly with the cold waters of the Benguela Current and sparsely vegetated desert on the west coast and render the land–sea thermal contrast less distinct for the east than for the west coast. Nevertheless, well-developed thermo-topographic boundary layer oscillations occur over Natal (Figure 2). Although topo-

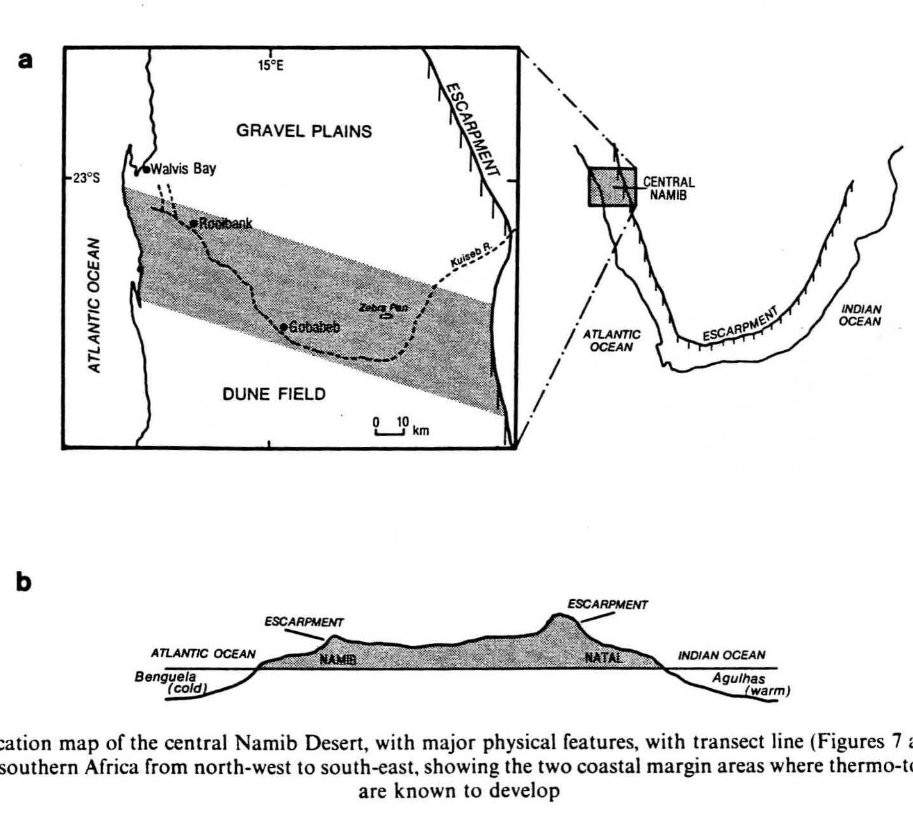

Figure 1. (a) Location map of the central Namib Desert, with major physical features, with transect line (Figures 7 and 8) shaded. (b) Section through southern Africa from north-west to south-east, showing the two coastal margin areas where thermo-topographic winds are known to develop

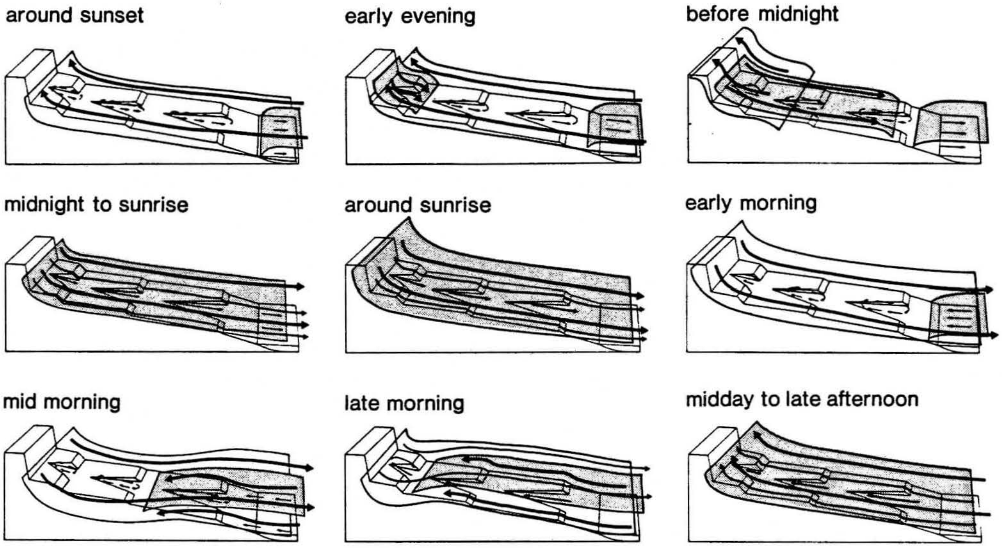

Figure 2. The diurnal variation of regional and local thermo-topographically induced winds between the eastern Escarpment, Natal, and the Indian Ocean (after Preston-Whyte and Tyson, 1988)

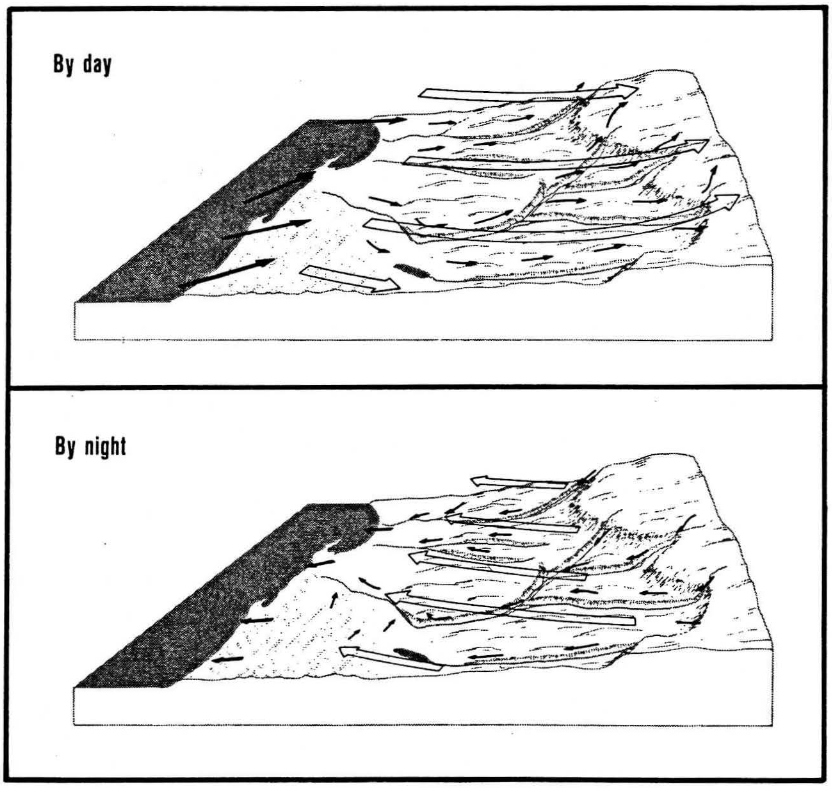

Figure 3. Schematic models of the local components of airflow over the central Namib Desert to show the occurrence of sea-breezes, valley winds and plain–mountain winds by day and in summer, and land-breezes, mountain winds, and mountain–plain winds by night in winter (after Tyson and Seely, 1980)

graphic influences on airflow over the west coast are expected to be less marked than those for the east coast, thermally induced boundary layer airflow should develop more strongly over the central Namib owing to the larger land–sea temperature contrast along the west coast. A conceptual model of local and regional diurnal airflow has been proposed for the eastern plateau slopes of Natal (Preston-Whyte and Tyson, 1988). Details are given in Figure 2; it remains to be seen to what extent a similar model will describe conditions in the central Namib.

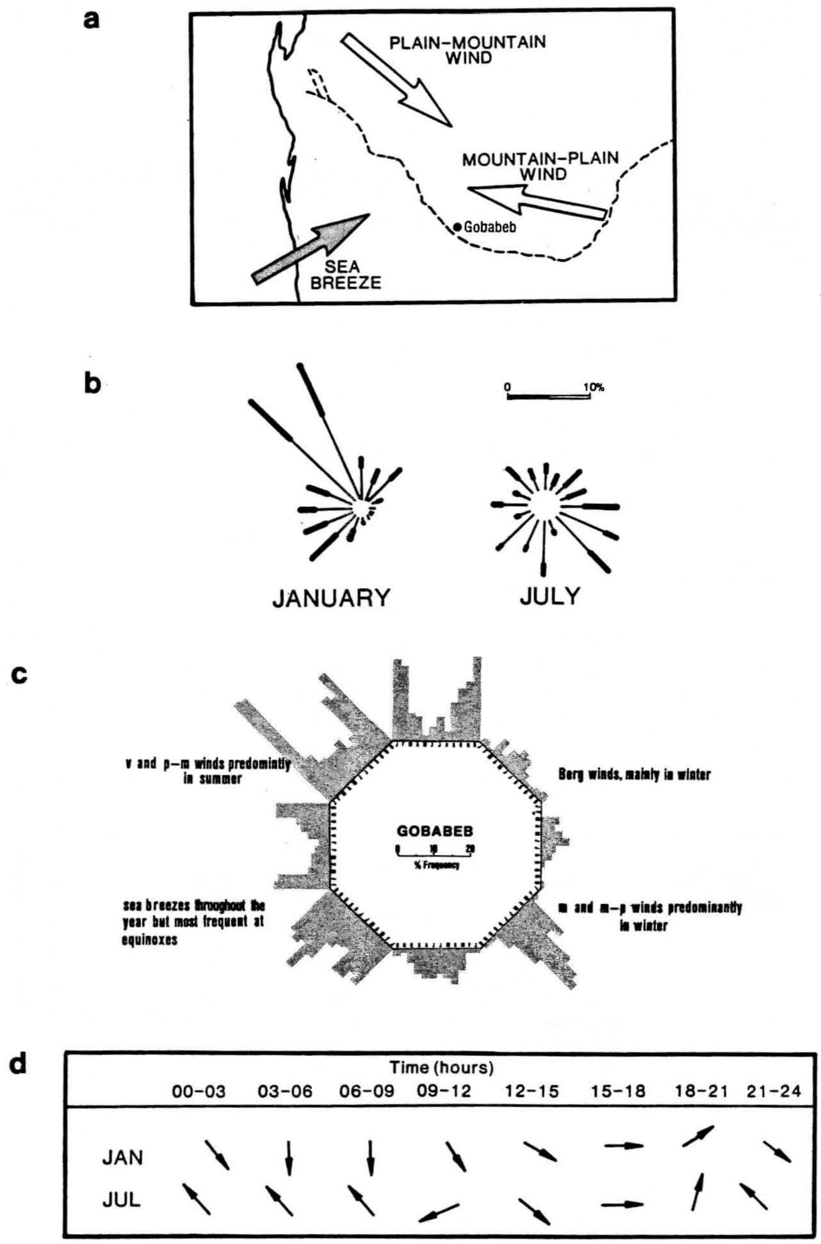

Figure 4. (a) Schematic diagram of the three major thermo-topographic winds prevailing in the near-surface circulation over the central Namib Desert. (b) Mean January and July surface wind-roses for Gobabeb for the period 1976–1981 (after Lancaster *et al.*, 1984). (c) Monthly variation of wind direction frequencies at Gobabeb. (d) January and July three-hourly surface wind vectors for Gobabeb for the period 1976–1981 (adapted after Lancaster *et al.*, 1984)

Surface boundary layer airflow over the central Namib is distinctive (Tyson and Seely, 1980) (Figure 3). Daytime south-westerly sea-breezes and north-westerly valley and plain–mountain winds dominate the near-surface circulation in summer, when calms are least frequent and regional pressure gradients reinforce the strong land–sea thermal contrast (Figure 4a–c). Nocturnal north-easterly land-breezes and south-easterly mountain and mountain–plain winds associated with cool air drainage occur preferentially in winter. Distinctive changes in nocturnal airflow patterns are evident at Gobabeb between summer and winter (Figure 4d). Interseasonal variations are less obvious during the afternoon–evening period (1200–2100 h) when surface heating effects are greatest. At times, synoptic disturbances may completely disrupt the diurnal rhythm of boundary layer airflow over the Namib, the most common such disturbance being the easterly to north-easterly Berg winds resulting from strong pressure gradients normal to the coast. Such disturbances are infrequent, however, and the surface climate of the central Namib is dominated by near-surface thermally and topographically induced boundary layer oscillations. A more detailed analysis of surface winds at Gobabeb confirms these findings (Lancaster *et al.*, 1984).

DATA

Hourly pilot balloon (PIBAL) ascents were made from Gobabeb in the central Namib (Figure 1a) during each of the months July 1986 and January 1988. On several suitable occasions, hourly ascents were also made from Rooibank, in the Kuiseb Valley 22 km from the coast, and at Zebra Pan, on the gravel plains north-east of Gobabeb and 100 km from the sea (Figure 1a). Simultaneous measurements enabled the construction of vertical sections of the boundary layer along a transect some 100 km long between the coast and the Escarpment, while continuous hourly PIBAL measurements at Gobabeb allowed the detailed investigation of the diurnal rhythms and interactions within the boundary layer over a point representative of much of the central Namib area. The programme of field observations provided in excess of 460 h of soundings during periods of generally light gradient flow.

Single-theodolite tracking of balloons was used, with cognizance being taken of the errors to which such tracking is susceptible (Reynolds, 1966; Boatman, 1974). Since no appropriate radiosonde data were available against which to verify the PIBAL data, double-theodolite tracking was used on several occasions during the summer period when the effects of surface heating on lapse rates were greater. Errors in heights and wind speeds were no greater than those found in similar studies elsewhere (e.g. Sturman and Tyson, 1981).

OBSERVATIONS

Plain–mountain winds

Surface anemometer observations at Gobabeb show clearly that north-westerly plain–mountain winds predominate in summer (Figure 4c) and are least frequent in winter. The plain–mountain winds appear to be purely antitriptic winds resulting from the thermal gradient established between the gravel plains to the north and the Escarpment to the east. During summer, when surface heating effects are at a maximum, the plain–mountain wind oscillates in depth and strength day after day, with little or no disturbance (Figure 5a, *lower*). At Gobabeb, maximum depths of 1000–1600 m and speeds of 10–15 m s^{-1} occur between 1600 h and 1800 h; minimum depths and speeds occur around and after midnight. Although the near-surface flow may be disturbed and may reverse around sunrise (with the occurrence of the reverse-direction mountain–plain wind), at levels near the top of the boundary layer the inland airflow may continue uninterruptedly for long periods.

In winter, when nocturnal cooling effects are strongest, the inland-directed plain–mountain wind is a daytime phenomenon only and is seldom as deep as its summertime counterpart (Figure 5a, *upper*). The contrast between the summer and winter examples of the plain–mountain wind over the central Namib is striking. In summer this wind is a constant determinant of the structure of the boundary layer; in winter the near-surface airflow is much more disturbed and many more local winds develop due to the clear diurnal reversals in low-level temperature gradients between the Escarpment and the coast and within the valleys of the western plateau slopes.

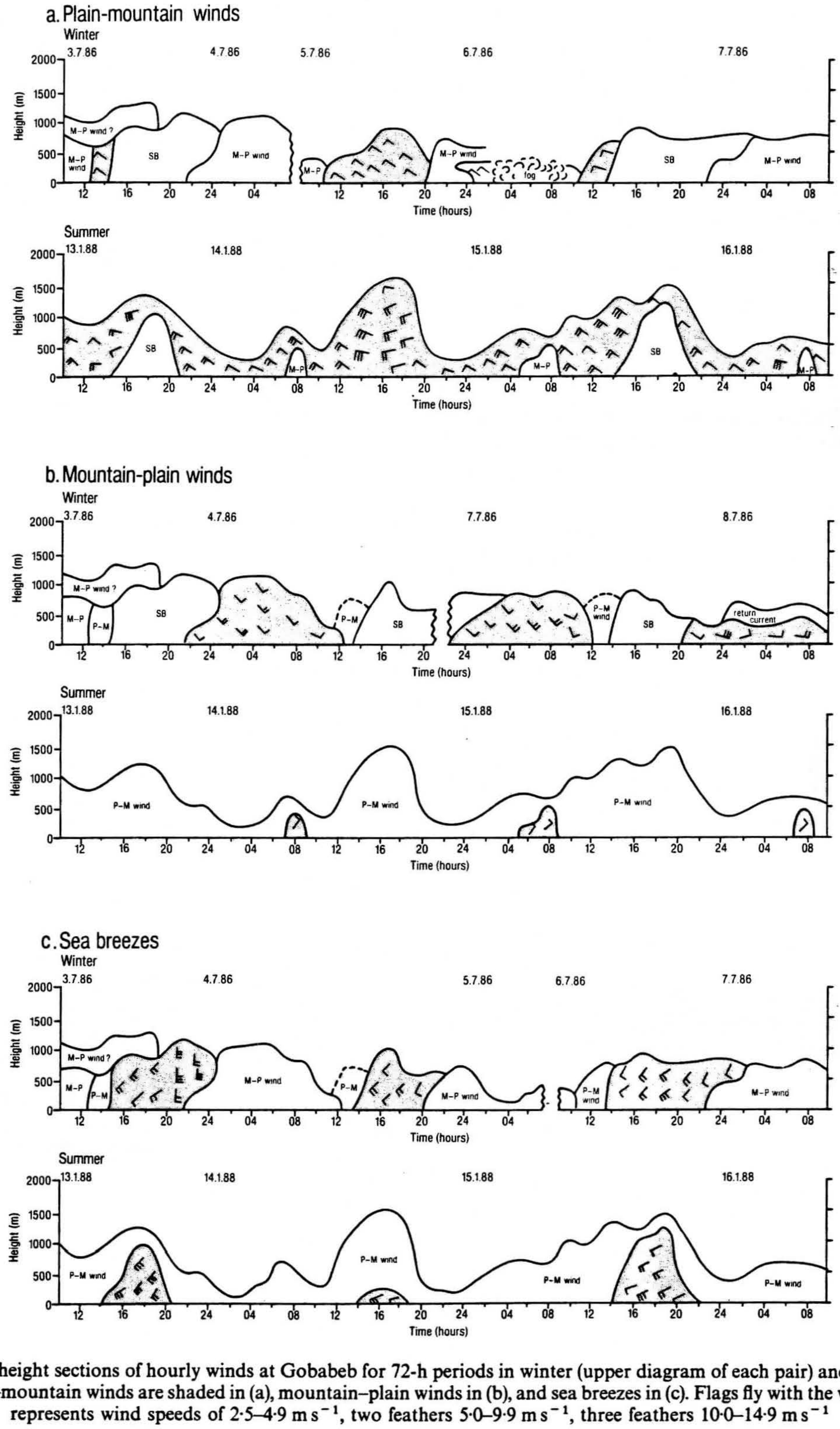

Figure 5. Time-height sections of hourly winds at Gobabeb for 72-h periods in winter (upper diagram of each pair) and summer (lower diagram). Plain–mountain winds are shaded in (a), mountain–plain winds in (b), and sea breezes in (c). Flags fly with the wind: one feather represents wind speeds of 2·5–4·9 m s^{-1}, two feathers 5·0–9·9 m s^{-1}, three feathers 10·0–14·9 m s^{-1}

Mountain–plain winds

Just as the boundary layer structure is dominated by plain–mountain winds in summer, so the winter pattern is characterized by regularly occurring nocturnal south-easterly mountain–plain winds blowing from the Escarpment zone towards the coast (Figure 5b, *upper*). These winds develop at Gobabeb around 2200 h, reach a maximum depth of about 1000 m and strength of 5–10 m s^{-1} after sunrise, and then decay rapidly between 1000 h and 1200 h the next day. This pattern is repeated day after day unless disturbed by synoptic-scale weather perturbations, such as Berg winds, coastal lows, or cold fronts. At times, when synoptic disturbances are entirely absent, the stable mountain–plain wind may develop a clear return current in a closed circulation seldom more than 1000 m deep. In summer, mountain–plain winds still occur, but only for a few hours at a time and seldom to a depth of even 500 m (Figure 5b, *lower*). It is surprising that they occur at all. The north-easterly direction of the summer mountain–plain winds reflects the inability of these poorly developed flows, unlike their winter counterparts, to appreciably modify the predominantly north-westerly gradient flow over the region.

Sea-breezes

Whereas the plain–mountain and mountain–plain winds over the central Namib are characterized by marked interseasonal differences in occurrence, sea-breezes occur regularly throughout the year, with a tendency to show a semi-annual cycle with peaks around the equinoxes (Figure 4c). The characteristic direction of airflow in the sea-breeze in the region is south-westerly, owing to turning of the breeze by the Coriolis effect (Jackson, 1954). On the Namib coast at Walvis Bay the sea-breeze is established between 0900 h and 1200 h on most mornings, attains maximum depths of about 1000 m and then decays until, after sunset, south-westerly winds are infrequent at the coast (Jackson, 1942, 1954). There is little seasonal variation in sea-breeze frequency at Walvis Bay, although Jackson (1954) suggests that the winter breeze is shallower than its summer counterpart. In winter the sea-breeze penetrates inland to reach Gobabeb around 1400 h, ceases at the ground at about 2000 h, when it is undercut by the developing mountain–plain wind, and continues blowing at heights of 500–750 m until around or after midnight (Figure 5c, *upper*). The system attains its maximum depth over the central Namib (seldom in excess of 1000 m) just before sunset. At the Gamsberg, 170 km inland, the sea-breeze reaches the Escarpment at about 1800 h and only lasts an hour or two before ceasing altogether.

In all the winter cases the evening sea-breeze at Gobabeb was undercut by the mountain–plain wind. In summer, when the breeze is not as strong as its winter counterpart and has a symmetrical diurnal pattern of growth and decay (Figure 5c, *lower*), this undercutting does not occur. The summer sea-breeze may exceed 1000 m in depth just before sunset, owing to the enhanced instability of the boundary layer. In summer the sea-breeze system over the central Namib decays from the top downward; in winter from the ground upward.

Wind profiles

Velocity profiles of the thermo-topographic winds induced by the thermal gradients and topography of the central Namib, and by the western and eastern boundary conditions imposed by the Benguela Current and the Escarpment, respectively, evidence a surprising regularity and symmetry (Figure 6). Return currents are difficult to measure, although on occasion such currents may be clear (as in Figure 6a). Jackson (1954) reports return currents several times the depth of the sea-breeze system at Walvis Bay, as does Atkinson (1981) for other parts of the world. At Gobabeb the return currents that were observed were always less deep than the sea-breezes below them. The nocturnal land-breeze is considerably weaker and shallower than its daytime sea-breeze counterpart (Figure 6b).

Plain–mountain winds in summer are deep and strong, and velocity maxima in excess of 10 m s^{-1} at heights of about 500 m are common (Figure 6c). The wintertime nocturnal mountain–plain winds (Figure 6d) are considerably weaker (with speeds of up to 3 m s^{-1}), but may reach similar depths to the plain–mountain winds. The zonal velocity maxima in both cases are attained at between 250 m and 400 m, that is approximately one-quarter to one-third the total depth of the system, although the maximum is less well defined for the mountain–plain than for the plain–mountain wind case. The sea-breeze reaches a maximum velocity normal to the coast at a height that is likewise one-quarter to one-third the depth of the system.

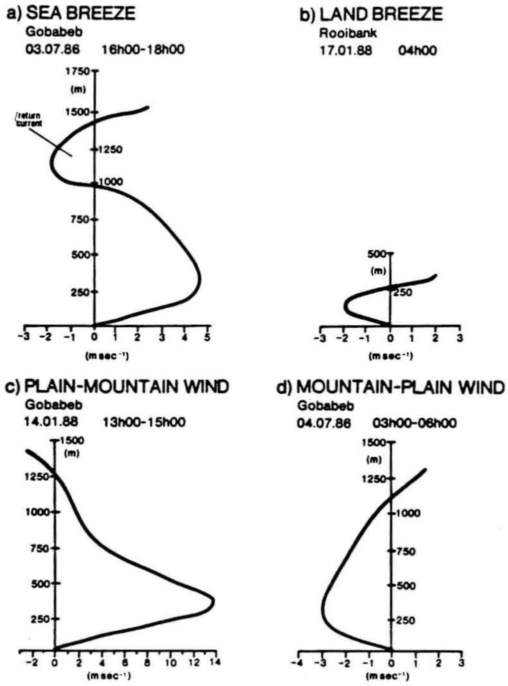

Figure 6. Vertical profiles of the zonal component (westerlies positive) of various thermo-topographic local winds at selected stations in the central Namib Desert

Airflow between the Escarpment and the coast

The vertical structure and temporal variation of the thermo-topographic boundary layer airflow over the central Namib have been described. Spatial continuity of this structure and of the oscillations of the boundary layer over time may be determined along transects on an approximately north-west–south-east section from seaward of Rooibank near the coast to east of Zebra Pan some 100 km inland (Figure 1a). Examples of these transects will be considered.

In summer a typical sequence of the decay of the plain–mountain wind and sea-breeze, followed by the onset of the oppositely directed mountain–plain wind, the decay of this wind and the re-establishment of the plain–mountain wind the following day, is given in Figure 7. A 1500-m-deep late-afternoon plain–mountain wind overlain by a sea-breeze some 500 m deep decayed rapidly after sunset, so that by 2100 h it was a skin of air with a depth of approximately 100 m moving slowly inland. The presence of a sea-breeze layer above the plain–mountain wind-flow between 1700 h and 1900 h appears unusual. From 2300 h until after 0100 h the inland penetration of the plain–mountain flow was prevented by drainage of cool air from the interior in the mountain–plain wind. At times this seaward drift of cool air was obliterated by a surge of warmer air moving inland; at times land-breezes developed temporarily on the coast. At 0600 h a thin layer of mountain–plain air draped the entire desert. Two hours later a plain–mountain wind had developed near the coast and penetrated inland beyond Gobabeb. By 1000 h the plain–mountain wind had advanced half-way toward the Escarpment.

A typical *winter sequence* shows the development of the same local winds as in summer, but with very different emphases (Figure 8). At 1600 h a strong sea-breeze (with a clear and spatially continuous return current) was displacing a decaying plain–mountain wind as it moved inland toward the Escarpment. The sea-breeze began weakening first at the coast; this weakening then progressed inland. By 1800 h, and for the following 2 h, the sea-breeze dominated the low-level airflow over the whole sea to Escarpment transect. By 2200 h a complete reversal had occurred and thereafter, throughout the night, drainage of cool air seaward dominated the local airflow regime to a depth of between 300 m and 500 m. At all times the local flow was

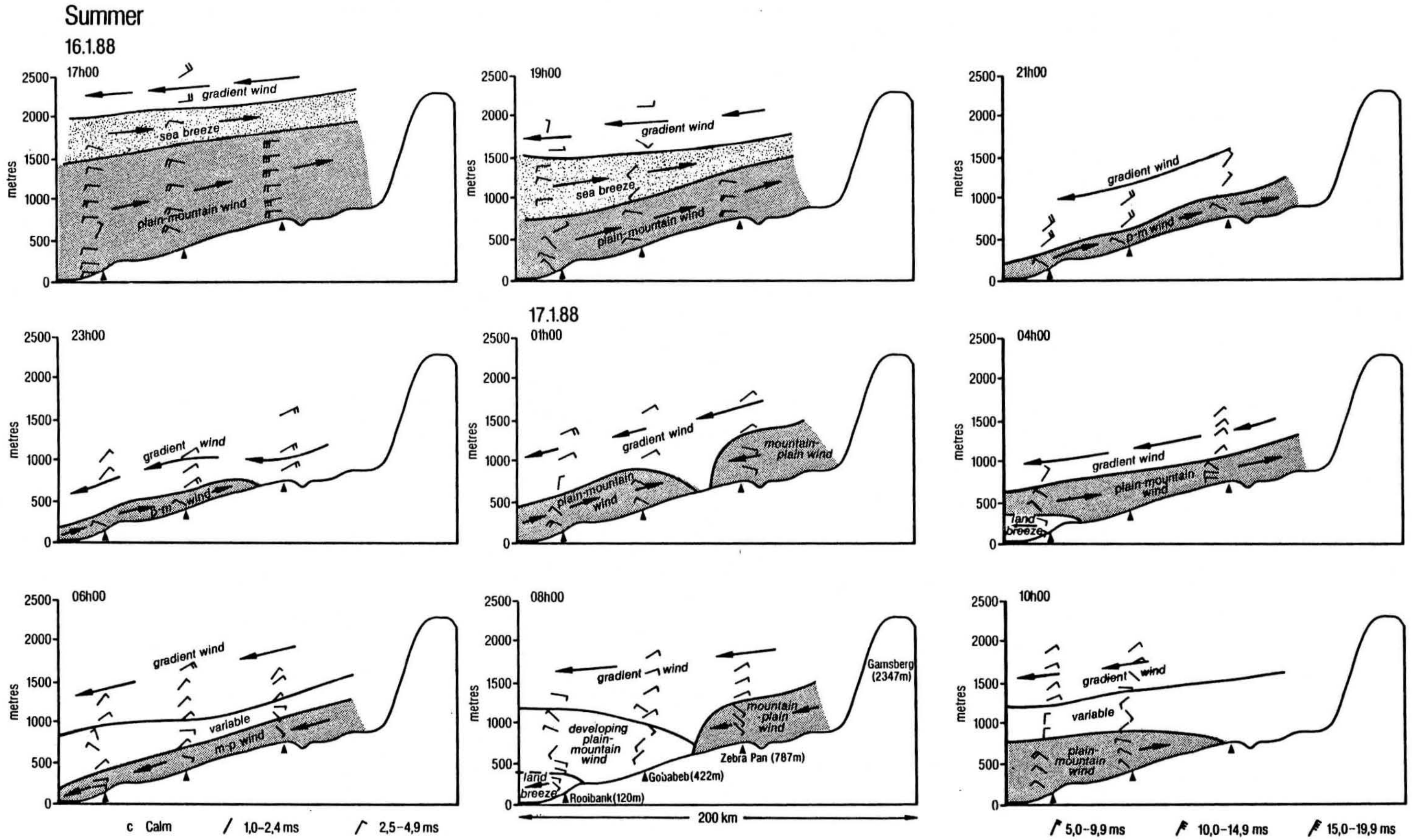

Figure 7. Time sequence of transects showing interactions of summertime thermo-topographic winds over the central Namib Desert between the coast and the Escarpment. Sea-breezes are stippled; mountain–plain and plain–mountain winds are shaded

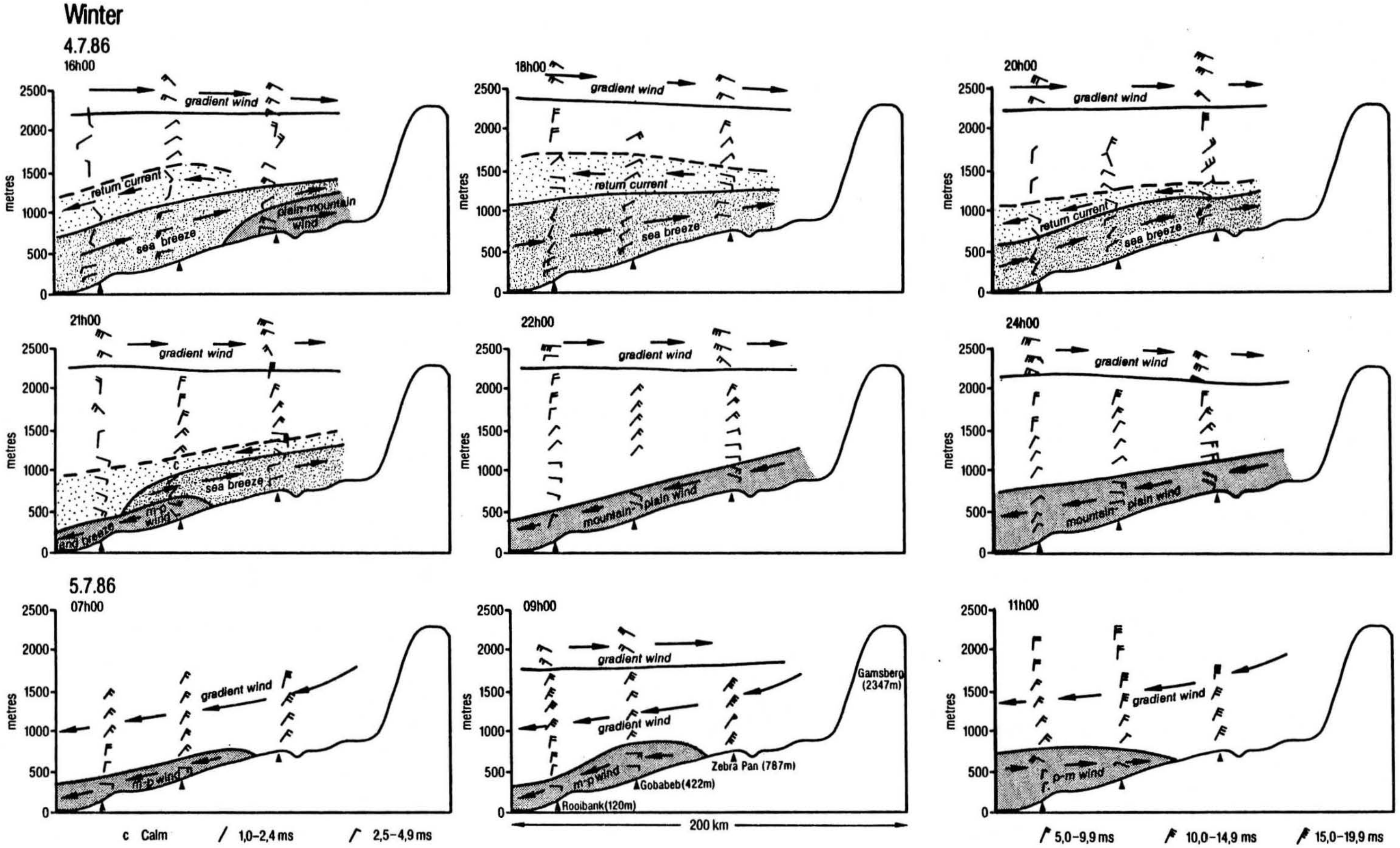

Figure 8. Time sequence of transects showing interactions of wintertime thermo-topographic winds over the central Namib Desert between the coast and the Escarpment. Sea-breezes and return currents are stippled; mountain–plain and plain–mountain winds are shaded

most prone to disturbance by synoptically driven winds, the further observations were made away from the Kuiseb River valley. Between 0900 h and 1100 h the following day the mountain–plain flow reversed abruptly to a plain–mountain wind, which then advanced inland as it strengthened throughout the day. The occurrence of two distinct layers of gradient flow on these days, with north-easterlies below the level of the Escarpment and north-westerlies above, reflects the presence of a shallow coastal low-pressure system centred north of Walvis Bay. Circulation around the southern part of the low resulted in the predominant north-easterly flow below about 2000 m, while the north-westerlies above that level conform to the prevailing large-scale synoptic situation.

DISCUSSION

The role of interacting thermo-topographic airflows in establishing the characteristics of the boundary layer over both the western (Tyson and Seely, 1980) and eastern (Preston-Whyte and Tyson, 1988) coastal margins of southern Africa has long been recognized. Over the central Namib, thermo-topographically induced boundary layer oscillations clearly dominate the diurnal and seasonal variations of local and regional airflow, confirming the identification of this area of southern Africa as one with particularly high diurnal variability of the near-surface windfield (Goldreich and Tyson, 1988). Disturbance of thermo-topographic airflow in the region by strong synoptic-scale winds is relatively rare (Tyson and Seely, 1980; Lancaster *et al.*, 1984) and the ratio of local to general winds gives it one of the highest meso-scale wind indices for the subcontinent (Goldreich and Tyson, 1988).

The particular characteristics of thermo-topographic regional airflow over the central Namib are the result of seasonally varying interactions among three major boundary layer wind systems: the sea/land-breeze response to the cold-ocean/hot-desert interface; the valley/mountain winds within the Kuiseb and similar river valleys dissecting the coastal plain; and the plain–mountain and mountain–plain winds resulting from the parallel existence of the desert plains and the Escarpment (Tyson and Seely 1980).

During undisturbed summer periods when surface heating effects over the central Namib are strongest, the warm, unstable north-westerly plain–mountain wind dominates the airflow within the boundary layer, both by day and by night, for days at a time. Its rhythm is broken only by short-lived punctuations near the ground by sea-breezes and mountain–plain winds. By contrast, the shallower east-coast plain–mountain winds, which reach depths of about 1200 m (Tyson, 1966, 1968b; Tyson and Preston-Whyte, 1972; Preston-Whyte and Tyson, 1988) compared with Namib plain–mountain winds more than 1500 m deep, are commonly replaced by nocturnal mountain–plain flow after midnight in both summer and winter.

The occurrence of cool, stable south-easterly mountain–plain winds over the central Namib is almost entirely limited to winter nights, when cooling effects are best developed. As is the case over Natal, west-coast mountain–plain winds are weaker and shallower than the plain–mountain winds and develop after sunset. The winter mountain–plain wind over the central Namib, however, persists for several hours longer than that over Natal (Tyson, 1968b; Tyson and Preston-Whyte, 1972; Preston-Whyte and Tyson, 1988). It may blow from an hour before sunset to late morning of the following day, is reasonably constant in depth, and evidences surging of the kind reported over Natal (Tyson, 1968c). Both north-westerly plain–mountain and south-easterly mountain–plain winds seldom bear any significant relationship to the near-surface pressure gradients of the general circulation. In their preferred seasons of development, diurnal valley and plain–mountain winds, and nocturnal mountain and mountain–plain airflows over the western desert coastal margins of southern Africa are better developed than similar regional airflows reported for the European Alps (Burger and Ekhart, 1937), the Rockies of North America (Hawkes, 1945), parts of Mexico (Lauer and Klaus, 1975), the Southern Alps of New Zealand (Sturman *et al.*, 1985) and the coastal and adjacent inland areas of Israel (Goldreich *et al.*, 1986).

Disruption of predominant plain–mountain winds over the central Namib by sea-breezes occurs briefly on most summer afternoons. During winter there is a clear diurnal oscillation of the boundary layer set up by the interaction between well-developed nocturnal mountain–plain winds, short-lived plain–mountain winds around midday, and the mid- to late-afternoon occurrence of the sea-breeze. Sea-breezes at Walvis Bay have been reported to occur with approximately equal frequency throughout the year, but to be most strongly developed during summer (Jackson, 1942). The latest sets of data for the central Namib show strongest sea-breezes in winter, a discrepancy attributable to the higher frequency of occurrence and dampening effect of middle- and high-level cloud on local wind development over the central Namib in summer. Sea-breezes, like all thermo-topographic winds, are best developed with completely clear skies. The suppressive effects of clouds on sea-breezes have recently been documented (Segal *et al.*, 1986) and may be considerable, although in New Zealand sea-breezes will develop in the presence of cloud (Sturman and Tyson, 1981).

The sea-breeze is stronger and deeper over the Namib coast than over the Natal shoreline (Jackson, 1954; Preston-Whyte, 1969; Preston-Whyte and Tyson, 1988); in both areas the characteristics of the breeze are in agreement with results obtained in both low latitudes (van Bemmelen, 1922; Roy, 1940; Dixit and Nicholson, 1964; Dekate, 1968) and middle latitudes (Frizzola and Fisher, 1963; Elliott, 1964; Gill, 1968; Johnson and O'Brien, 1973; Camuffo *et al.*, 1979; Bitan, 1981; Sturman and Tyson, 1981; Mathews, 1982; McKendry *et al.*, 1986; Prezerakos, 1986; Helmis *et al.*, 1987). After setting in at the coast at about 0900 h, the Namib sea-breeze backs to south-west through the day under the influence of the Coriolis force. Similar turning of the sea-breeze has been reported for other areas (Haurwitz, 1947; Staley, 1957; Estoque, 1961, 1962; Frizzola and Fisher, 1963; Hsu, 1970; Yan and Anthes, 1987). By mid-afternoon the sea-breeze is south-westerly throughout the boundary layer from the coast to beyond Gobabeb, and by sunset the sea-breeze has penetrated almost to the Escarpment. Weakening of the sea-breeze at the coast around sunset coincides with maximum south-westerly flow at Gobabeb, is followed several hours later by a sea-breeze maximum further inland and then by penetration of the wind to the Gamsberg on the plateau an hour or so before midnight. The sea-breeze dominates boundary layer airflow over the entire central Namib between the coast and the Escarpment for much of the afternoon and early evening period, giving way around midnight at Gobabeb to relatively stable mountain–plain flow advancing seaward from the Escarpment, although the breeze has ceased at the coast several hours before. Unlike its plain–mountain and mountain–plain counterparts, the sea-breeze may on occasion be associated with a clear return current.

In contrast to the deep, unstable afternoon sea-breeze circulation, the nocturnal land-breeze of the coastal littoral is a shallow, short-lived system. Over the east coast of southern Africa, the land-breeze is seldom deeper than 300 m (Preston-Whyte, 1968, 1974); that on the west coast is similar in depth. Onset of land-breeze flow on the southern African west coast after 0100 h and cessation of the breeze before 0900 h accords with observations of land-breezes in India (Dekate, 1968; Aggarwal *et al.*, 1980) and North America (Meyer, 1971; Atkinson, 1981), and with theoretical studies (Rotunno, 1983).

The vertical structures of west-coast boundary layer airflow across the central Namib are similar to those of east-coast Natal thermo-topographic winds (Preston-Whyte and Tyson, 1988). Lagrangian profiles of the wind component normal to each of the coasts show that the maximum velocities of land- and sea-breezes, plain–mountain, and mountain–plain winds are comparable for both regions, although west-coast airflow systems are deeper, particularly in summer. Over the west coast, as in Natal (Tyson, 1966, 1968a, b, c; Tyson and Preston-Whyte, 1972; Preston-Whyte and Tyson, 1988), maximum-velocity airflow occurs at 250–300 m above the surface. The variation of wind speed with height may be modelled using a parabolic profile in which the height of maximum velocity is half the height of the system (Davidson and Rao, 1963), or using a Prandtl profile in which the height of maximum velocity is one-quarter the height of the system (Defant, 1958; Atkinson, 1983; Pielke, 1984). The parabolic profile gives a good representation of mountain, mountain–plain, and plain–mountain winds in the Drakensberg region, and of the Natal-coast sea-breeze (Preston-Whyte and Tyson, 1988), whereas the Prandtl profile best fits mountain winds in the Natal interior (Tyson, 1968c) and Drakensberg valley winds. Central Namib wind profiles of plain–mountain and mountain–plain winds and of the sea-breeze, by contrast, conform closely to the Prandtl model. The plain–mountain wind maximum occurs at a height of ± 300 m in a system some 1200 m deep, and the sea-breeze maximum at 250 m in a 1000 m-deep system. The west-coast land-breeze, however, exhibits a parabolic profile, with maximum wind speeds at approximately half the depth of the system.

CONCLUSIONS

In earlier studies the nature of the thermo-topographic forcing of boundary layer airflow over the eastern plateau slopes of southern Africa between the Drakensberg and the Natal coast was clearly demonstrated. On the coast, land- and sea-breezes were shown to interact temporally and spatially with mountain and valley winds of the adjacent dissected coastal hinterland. Further inland, within the deep valleys of Natal, the occurrence and effect of strong, deep mountain and valley winds were shown to dominate airflow in the lower boundary layer. Between the Escarpment and the sea, regional mountain–plain and plain–mountain winds were demonstrated to occur regularly. Over Natal the integrated effect of these three thermo-topographic wind systems is to produce deep and regular meso-scale influxes of air toward the Escarpment by day and oppositely directed effluxes of air by night in a distinctive oscillation of the boundary layer. The same has now been shown to occur on the west coast on the opposite side of the subcontinent.

The strong thermal gradients across the central Namib Desert, bounded to the west by the consistently cold ocean and to the east by the Escarpment, strongly heated by day and similarly cooled by night, together with the pronounced diurnal heating and cooling of the desert itself, are responsible for the strength and longevity of the boundary layer airflows developed over the region. Throughout the year the boundary layer varies in depth from under 500 m at night to over 1000 m by day, as a pulsating mass of air moves alternately inland and thereafter seaward in an oscillation as distinctive as that reported anywhere. The seasonally clearest suite of successional winds is encountered in winter when the plain–mountain winds are weak and short-lived and the boundary layer is dominated by mountain–plain winds by night and sea-breezes during the post-noon period. In summer, north-westerly plain–mountain winds are present by day and night in a remarkably persistent response to thermal and topographic forcing. These distinctive boundary layer oscillations are disrupted only infrequently by strong synoptic-scale disturbances. Surface characteristics and vertical structure of the lower atmosphere over the central Namib confirm that, in the region between the coast and the inland plateau, boundary layer airflow is controlled by surface thermal effects and by both local and regional topography. Thermo-topographic airflows in the region frequently have a regional significance equalling or exceeding that of the general circulation. By virtue of the strength, depth, and unusually clearly defined diurnal and seasonal oscillations of the thermo-topographic airflows of the region, the central Namib Desert constitutes a unique laboratory for the study of boundary layer oscillations.

ACKNOWLEDGEMENTS

The staff of the Desert Ecological Research Unit at Gobabeb, particularly Dr M. K. Seely, provided invaluable facilities, assistance, and hospitality. Their help is gratefully acknowledged. The study would not have been possible without the assistance of staff and students of the University of the Witwatersrand. Our thanks go to Dr M. J. Wilkinson and Mrs C. H. Vogel for helping to organize and run the field trips. Kathy Barbafiera, Mario Barbafiera, Brynn Bayman, Brenda Beverly, Kim Christie, Lynne Collie, Sue Crouch, Caryn Eccles, Gavin Elliot-Wilson, Gordon Fenwick, Diane Geer, Jenny Gobey, Michael Haynes, Yvonne Hong, Greg Knill, Jerry Lengoasa, Paul Maclear, Christine Muller, Charles Nepgen, Sean O'Beirne, Michelle Pearse, Colin Pilkington, Mike Slattery, Belinda Theron, Greg Theron, Jenny Tyson, Anja van Nierop, and Ingo Vogt assisted with data collection. They constituted a marvellous team, and without their assistance the study could not have proceeded. Mr P. Stickler drew the diagrams. The research was funded in part through the Foundation for Research Development of CSIR.

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