The influence of a synoptic-scale disturbance on topographically induced boundary layer circulations over the central Namib Desert

J.R. LENGOASA, J.A. LINDESAY AND A.M. VAN NIEROP

Climatology Research Group, University of the Witwatersrand, Johannesburg 2050, South Africa

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ABSTRACT

Local climates of coastal areas are characterised by the modification of synoptic-scale airflow by thermally and topographically induced diurnal oscillations of the lower boundary layer. At times, however, synoptic-scale disturbances may completely disrupt the diurnal rhythm of boundary layer airflow. Hourly pilot balloon ascents made at three locations in the central Namib before, during and after the passage of a low-pressure disturbance are used to show the extent of the disruption of boundary layer airflow patterns by the disturbance, and to illustrate the importance of such disturbances for low-level winds between the coast and the Escarpment. The complex interactions of circulations in low and middle latitudes with the disturbance influence both the intensity and behaviour of the coastal system.

INTRODUCTION

Local climates of coastal and mountain areas are characterised by the modification of synoptic-scale airflow by thermally and topographically induced diurnal oscillations of the lower boundary layer. Individual thermo-topographic wind systems, e.g. mountain/plain winds and land/sea breezes, have been studied in many parts of the world (Defant 1958, Flohn 1969, Atkinson 1981, Yoshino 1981, Sturman 1987), and in southern Africa the interactions of boundary layer airflows have also received attention (e.g. Tyson 1964, 1986, Preston-Whyte 1974, Tyson and Seely 1980, Jury *et al.* 1990, Lindesay and Tyson 1990). Much of the work on boundary layer airflows in southern Africa has been concentrated on the eastern Escarpment and coastal margins of the south-eastern subcontinent (reviewed in Tyson 1986). More recently, however, an analysis of 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 over the central Namib Desert has been attempted by Lindesay and Tyson (1990).

The physical setting of the central Namib Desert, bounded on the west by the cold Atlantic Ocean and to the east by the Escarpment, is not only important to this study because it creates favourable conditions for a distinctive sequence of diurnal and seasonal wind patterns, but also because it is conducive to the formation of coastal disturbances such as coastal lows. Although much work has been done on the characteristics and effects of the coastal low at the land-sea interface (Taljaard *et al.* 1961, Preston-Whyte 1975, Diab 1984, Hunter 1984, Jury 1984, Kamstra 1984, Nelson and Glyn-Thomas 1984, Sciocatti 1984, Shillington 1984, Walker 1984, Heydenrych 1987, Jury *et al.* 1990, Reason and Jury 1990, Reason and Steyn 1990), the effects of these and similar disturbances as far north as the central Namib and inland from the coast remain undocumented. A workshop dealing specifically with coastal lows (Coastal Low Workshop 1984) highlighted the value of understanding the distinctive disruption of the regular coastal weather pattern caused by these disturbances. Most importantly, coastal lows are accompanied by sharp changes in weather conditions including changes in wind direction and speed, temperature, pressure and the height of the non-surface inversion.

In this paper synoptic-scale disruption of the diurnal oscillations of the boundary layer over the central Namib by the passage of a low-pressure disturbance will be examined. The interaction of the coastal disturbance with other features of the synoptic-scale circulation, including an easterly wave, will be discussed. Both the vertical and horizontal structures of the thermo-topographic airflow, prior to, during and after the passage of the disturbance will be considered, and the extent of the disruption of normal airflow patterns evaluated.

DATA AND METHODS

Hourly pilot balloon (pibal) ascents were made from Gobabeb in the central Namib (Fig. 1) during January 1988, including the period 11-13 January during the passage of a synoptic-scale disturbance. On several occasions hourly ascents were also made at Rooibank and Zebra Pan. Simultaneous measurements enabled the construction of vertical sections of the boundary layer along

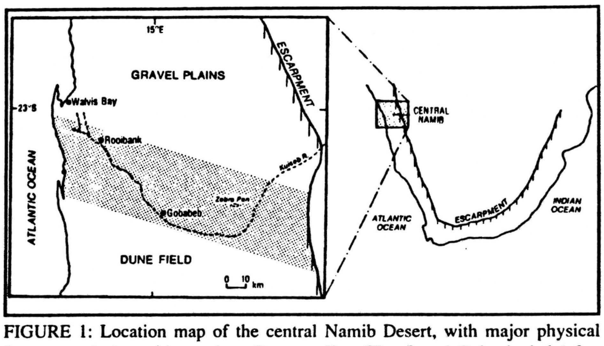

FIGURE 1: Location map of the central Namib Desert, with major physical features and observation points. Transect line (Figs 3 and 6) is shaded (after Lindesay and Tyson 1990).

a 100 km-long transect between the Escarpment and the coast, and continuous pibal ascents at Gobabeb allowed the detailed investigation of the diurnal rhythms and interactions within the boundary layer.

PHYSICAL SETTING AND GENERAL AIRFLOW CLIMATOLOGY

The central Namib Desert on the west coast of southern Africa is bounded by the cold Atlantic Ocean to the west and a dissected plateau slope beneath the Escarpment some 160-180 km to the east (Taljaard 1979, Tyson and Seely 1980, Lancaster *et al.* 1984) (Fig. 1). The deeply incised Kuiseb River valley forms a boundary between the relatively flat gravel plains to the north and linear dunes of the sand sea to the south. Large thermally induced boundary layer airflows develop strongly over the central Namib owing to the large land-sea temperature contrast along the west coast (Munn 1966, Flohn 1969). Surface boundary layer airflow over the central Namib is distinctive (Tyson and Seely 1980). The predominant summer winds are the south-westerly sea breeze and north-westerly plain-mountain winds, when calms are least frequent and regional pressure gradients reinforce the land-sea thermal contrast. Sea breezes occur throughout the year with a tendency to show a semi-annual cycle. The characteristic direction of airflow in the sea breeze in the region is south-westerly, owing to the turning effect of the coriolis force (Jackson 1954). In summer the sea breeze is characteristically not strong and has a symmetrical diurnal pattern of growth and decay (Lindesay and Tyson 1990) (Fig. 2a). The sea breeze may exceed 1000 m in depth just before sunset owing to the enhanced instability of the boundary layer. The plain-mountain winds are dependent on 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 (Fig. 2b). A typical summer sequence of boundary layer winds includes the growth and decay of the plain-mountain wind and sea breeze, followed by the onset of the oppositely directed mountain-plain wind, (Fig. 3), the decay of this wind and the re-establishment of the plain-mountain wind the following day.

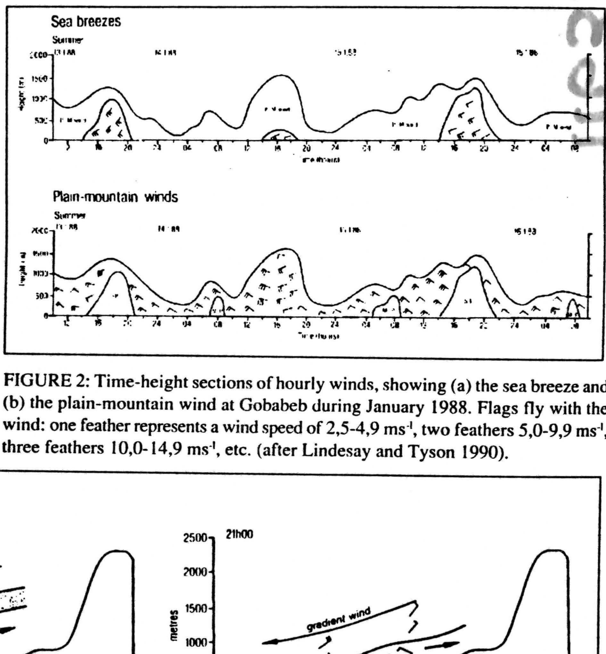

FIGURE 2: Time-height sections of hourly winds, showing (a) the sea breeze and (b) the plain-mountain wind at Gobabeb during January 1988. Flags fly with the wind: one feather represents a wind speed of 2,5-4,9 ms^{-1}, two feathers 5,0-9,9 ms^{-1}, three feathers 10,0-14,9 ms^{-1}, etc. (after Lindesay and Tyson 1990).

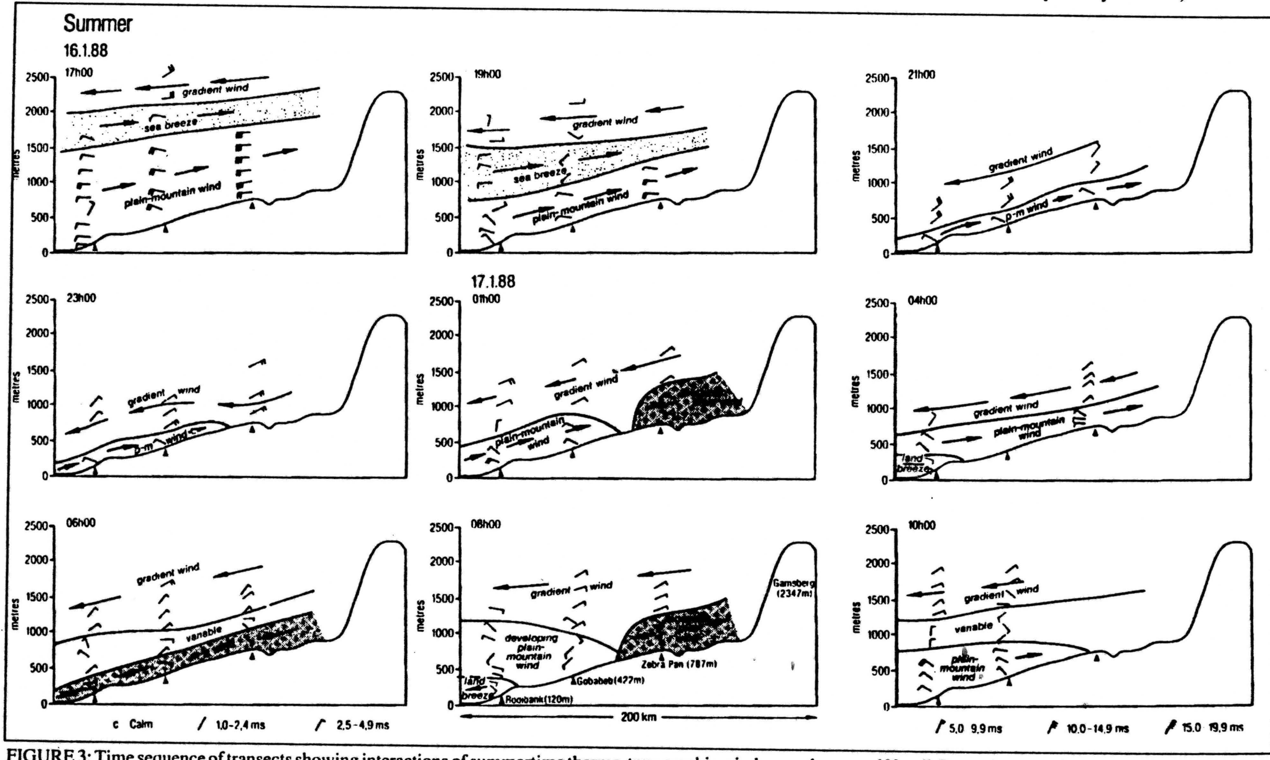

FIGURE 3: Time sequence of transects showing interactions of summertime thermo-topographic winds over the central Namib Desert between the coast and the Escarpment. Sea breezes are stippled; mountain-plain and plain-mountain winds are shaded (after Lindesay and Tyson 1990).

WEST COAST DISTURBANCES

The generalised boundary layer airflow patterns over the central Namib are occasionally disrupted by synoptic-scale disturbances such as the berg wind, which is most common in winter, and the coastal low which is more commonly a summer phenomenon. The term coastal low is reserved for a unique low pressure system which develops along the coast, is generally shallow, and does not extend into the atmosphere to a depth greater than the elevation of the plateau (Estie 1984, Reason and Steyn 1990). Over the central Namib this definition excludes low pressure systems which lie within the trough in the easterlies which frequently extends over this region in summer and lies above the Escarpment.

Two distinct synoptic situations which encourage the formation of coastal lows and similar systems have been identified. The first occurs when a high pressure cell ridges in to the south of the continent, causing offshore flow over the west coast, and the second is characterised by an approaching depression or trough in the westerlies resulting in offshore flow over the south coast (De Wet 1984, Estie 1984, Walker 1984). As the central Namib is generally influenced by the easterlies rather than the westerlies, the discussion in this paper will be limited to the first synoptic type. These particular coastal lows are referred to as the 'summer west coast low' (Coastal Low Workshop 1984), occurring most frequently between October and March (Walker 1984, Heydenrych 1987).

Two models equating the structure of southern African coastal lows with that of coastally-trapped Kelvin waves in the ocean and atmosphere have been proposed (Gill 1977, Nguyen and Gill 1981, Reason and Steyn, 1990). Due to the disturbance intensity decreasing with distance from the coast it has been concluded that the low-level non-surface inversion (Taljaard 1955, Diab, 1986) and the Escarpment provide conditions favourable for the trapping of disturbance energy both vertically and laterally. The trapped disturbances propagate horizontally along the coast, and are modified by the changing forcing function. This model accounts for both the migratory and dissipative characteristics of coastal lows. As a coastal low passes any point inland from the coast, the wind is expected to veer from north-easterly, easterly or south-easterly to north-westerly, westerly or south-westerly (Fig. 4). The abrupt change in wind direction is regarded as a clear indication of the passage of a coastal low (Estie 1984, Jury 1984, Walker 1984), and is distinct from the backing of the wind which is a normal characteristic of the behaviour of the surface wind with the development of a sea breeze. Available boundary layer wind data are therefore used in this study as an indicator of the presence and inland extent of the influence of a summer low-pressure disturbance, and to show the disruption of the normal thermo-topographic airflow pattern by the low and associated synoptic circulation features.

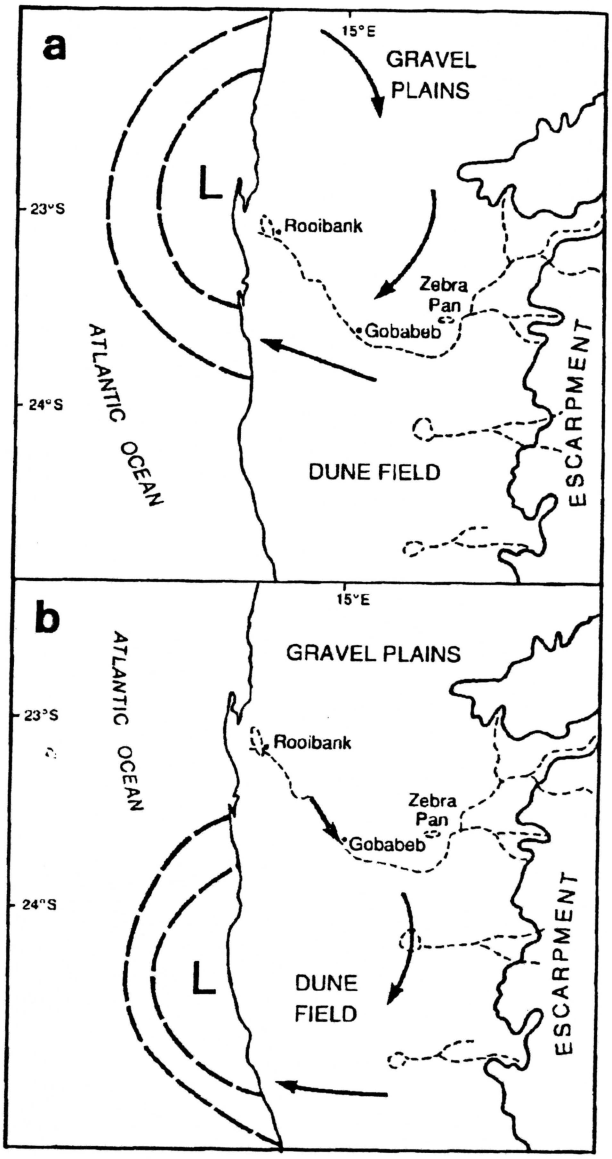

FIGURE 4: Schematic model of the expected progression of winds with the southward movement of a coastal low pressure system along the west coast of southern Africa.

OBSERVATIONS

Necessary conditions for the formation of coastal lows and other coastal disturbances are the high interior plateau and the presence of off-shore gradient winds. Data recorded over the study period give strong empirical evidence of the association between offshore flow over the Escarpment and coastal disturbance occurrence. On 11 January 1988 (Fig. 5a) the gradient wind over the central Namib was generally south-easterly, overlying changing boundary layer airflow. A north-westerly plain-mountain wind was present around 10:00, and was replaced by a south-westerly sea breeze from 12:00. Flow in the boundary layer then backed further to become south-easterly following the decay of the sea breeze around 22:00 (Fig. 5a). The backing of the wind is a common characteristic of the behaviour of near-surface airflow in the presence of a sea breeze. The gradient wind became more directly offshore (i.e. easterly) with higher wind speeds on 12 January 1988 (Fig. 5b). Strong onshore westerly to south-westerly flow below 1000 m after midday indicated the presence of a synoptic-scale disturbance. The south-westerly flow was first observed at Gobabeb at 11:00, some hours before the sea breeze would be expected there. The horizontal profiles (Fig. 6) show an offshore gradient wind at Rooibank at 17:00 on 12 January 1988, when the effects of the disturbance at Gobabeb were clearly evident in south-westerly flow and the plain-mountain wind was still dominant at Zebra Pan. By 18:00, however, flow was south-westerly throughout the boundary layer as far inland as Zebra Pan, and velocities reached 20 ms^{-1} over Gobabeb. The disturbance had intensified by 19:00 and was particularly strong over Gobabeb. Thunderstorm activity was evident along the Escarpment around this time. Observations were impossible after 23:00 on 12 January 1988 (Fig. 5b) due to high wind speeds and turbulence in the boundary layer, and were resumed only after 06:00 on 13 January 1988 (Fig. 5c) when the normal morning wind pattern of north-westerly to westerly plain-mountain flow was re-established. This was punctuated by the sea breeze between 15:00 and 20:00, while gradient flow remained north-easterly, i.e. offshore (Fig. 6). Clearly, the effects of the disturbance extended at least as far inland as Zebra Pan, some 100 km from the coast.

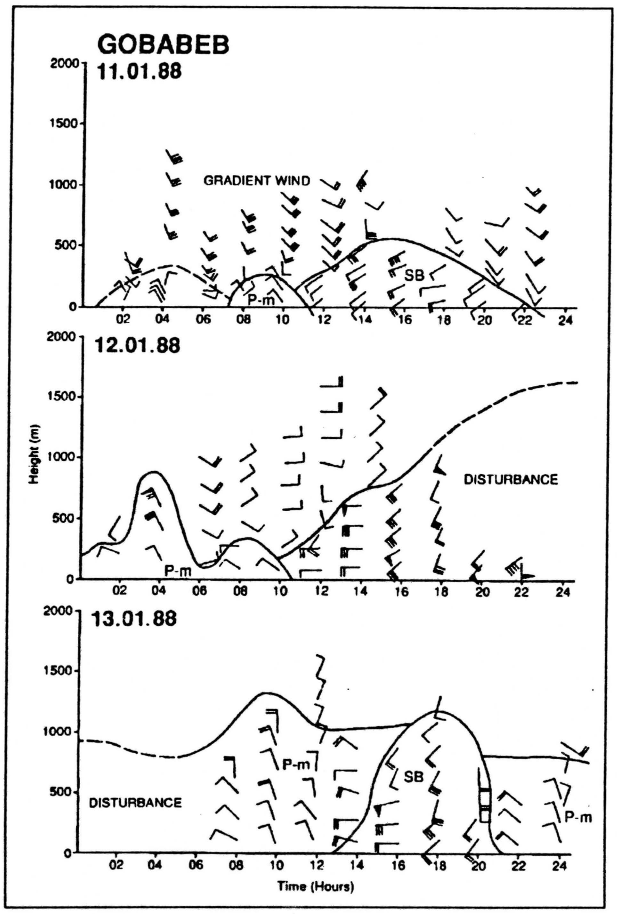

FIGURE 5: Time-height sections of hourly winds at Gobabeb showing the normal progression of winds during a 24-hour period (11 January 1988), and the effect of the disturbance (12-13 January). Flags fly with the wind: one feather represents a wind speed of 2,5-4,9 ms^{-1}, two feathers 5,0-9,9 ms^{-1}, three feathers 10,0-14,9 ms^{-1}, triangles 20,0-24,9ms^{-1}.

The strength of the disruption of the boundary layer wind field induced by the presence of the disturbance is illustrated by vertical profiles of the zonal component of the wind (Fig. 7). Under undisturbed conditions, regularity and symmetry characterise the velocity profiles of the thermo-topographic winds over the central Namib, and most local winds exhibit parabolic profiles (Lindesay and Tyson 1990). Two boundary layer winds, the sea breeze and the plain-mountain wind, have predominantly westerly zonal components. Westerly flow extends from the surface up to approximately 1000 m, with wind speeds as high as 5 ms^{-1} in the case of the sea breeze and speeds up to 14 ms^{-1} in the plain-mountain wind (Lindesay and Tyson 1990). On 12 January 1988 winds in the boundary layer were predominantly westerly (onshore), but the

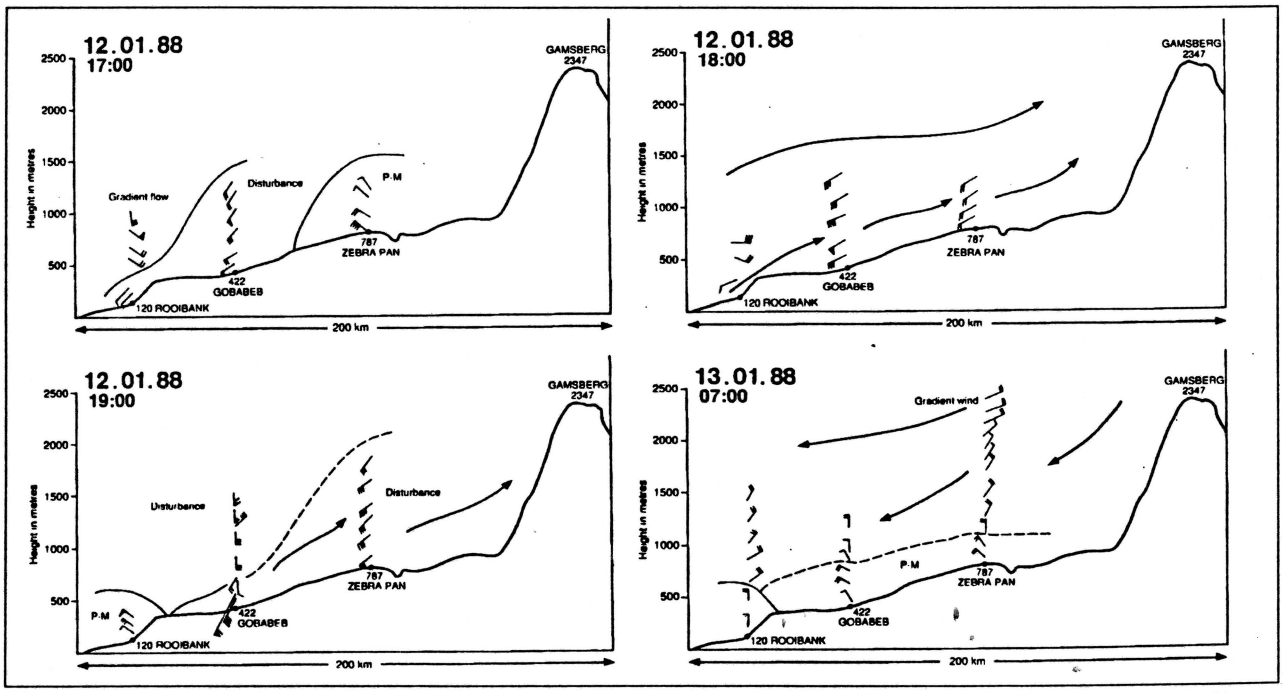

FIGURE 6: Time sequence of transects showing the depth and extent of the disturbance (12 January 1988), and re-establishment of the normal summertime pattern of thermo-topographic winds (13 January 1988). Flags fly with the wind: one feather represents a wind speed of 2,5-4,9 ms^{-1}, two feathers 5,0-9,9 ms^{-1}, three feathers 10,0-14,9 ms^{-1}, triangles 20,0-24,9 ms^{-1}, etc.

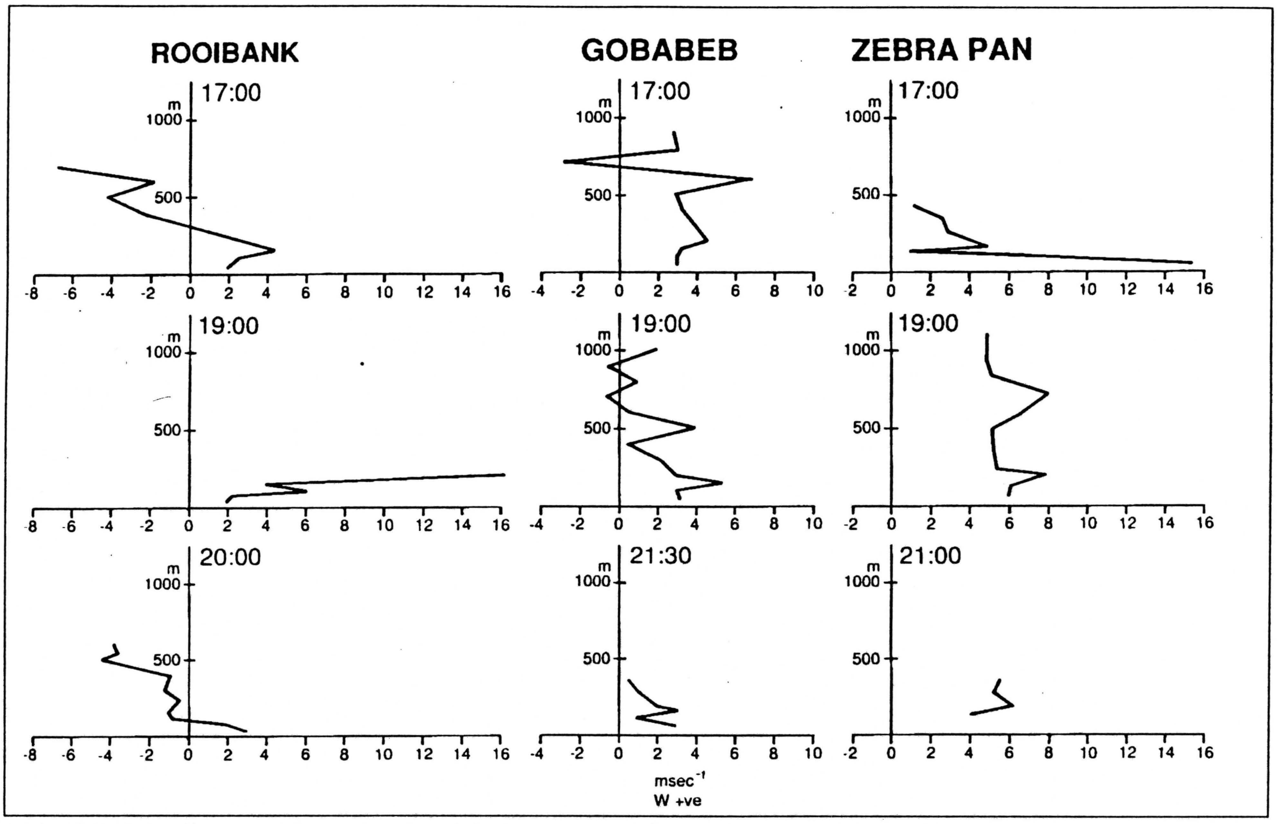

FIGURE 7: Vertical profiles of the zonal component of the wind at three central Namib Desert stations at selected hours on 12 January 1988, during the presence of the low. Westerly components are positive.

vertical profiles of the zonal components (Fig. 7) were quite different from the sea breeze and plain-mountain wind profiles. Wind speeds varied considerably with height, with occasional strong gusts and no clearly defined pattern of growth/decay of the westerly component with time. Turbulent disruption of the normal boundary layer flow by the disturbance is clearly evident.

DISCUSSION

The characteristic distinct disruption of the normal boundary layer wind patterns in the central Namib by a synoptic-scale disturbance is clearly illustrated by observations at Gobabeb, Rooibank and Zebra Pan between 11 and 13 January 1988. The strength of the disruptive effects of the disturbance at Gobabeb (approximately 56 km inland) and Zebra Pan (approximately halfway between the coast and Escarpment, the Escarpment marking the inland limit of possible influence) suggests that the disturbance was not simply a coastal low but was enhanced in some way not evident from the available ground observations. The early onset of south-westerly flow at Gobabeb is an indicator of the initial effects of the disturbance, as south-westerly winds associated with sea breezes in the central Namib are a mid-afternoon feature of the local circulation. Thus the time of onset as well as the depth of this particular system rule out the possibility of this being a sea breeze rather than a synoptic-scale disturbance. The strength of the winds with the onset of the south-westerlies is comparable with the 'busters' associated with coastal lows along the Natal coastline (Preston-Whyte 1975, Diab 1986), but the evolution of the windfield is not consistent with the usual pattern of coastal low development.

Analysis of observations at Rooibank, Gobabeb and Zebra Pan shows that the disruption of boundary-layer winds that characterised the presence of the disturbance at the inland stations is not as evident close to the coast. At Rooibank winds were not disrupted to the same vertical extent (Fig. 6) or for as long as at the other stations. This could be attributed to the location of Rooibank in a shallow valley which offers some protection from gradient flow and mesoscale circulation conditions, whereas Gobabeb and Zebra Pan are in more exposed positions. The wind changes at Zebra Pan, approximately 100 km inland, on 12 January 1988 support the idea that the disturbance penetrated inland at least halfway to the Escarpment. The effects of the disturbance were apparent at Zebra Pan some two hours later than at Gobabeb, probably because Zebra Pan is approximately 50 km further inland, indicating that the disturbance originated at or near the coast.

Despite the observational evidence that the disturbance of 11-13 January 1988 did not exhibit characteristics of a typical coastal low, the standard available synoptic information indicates only the conditions that should produce such a feature (Fig. 8). The surface synoptic charts show the passage of two midlatitude depressions and the associated displacement of the South Atlantic Anticyclone, accompanied by onshore airflow and the formation of the coastal low. No defined troughs are evident in the tropical easterly flow. At 700 hPa (above the level of the plateau) the circulation appears to be anticyclonic and divergent, with offshore airflow. Nevertheless the boundary layer windfield did not display the adjustments usually associated with the passage of a coastal low. On the west coast winds are expected to change from southerly to south-

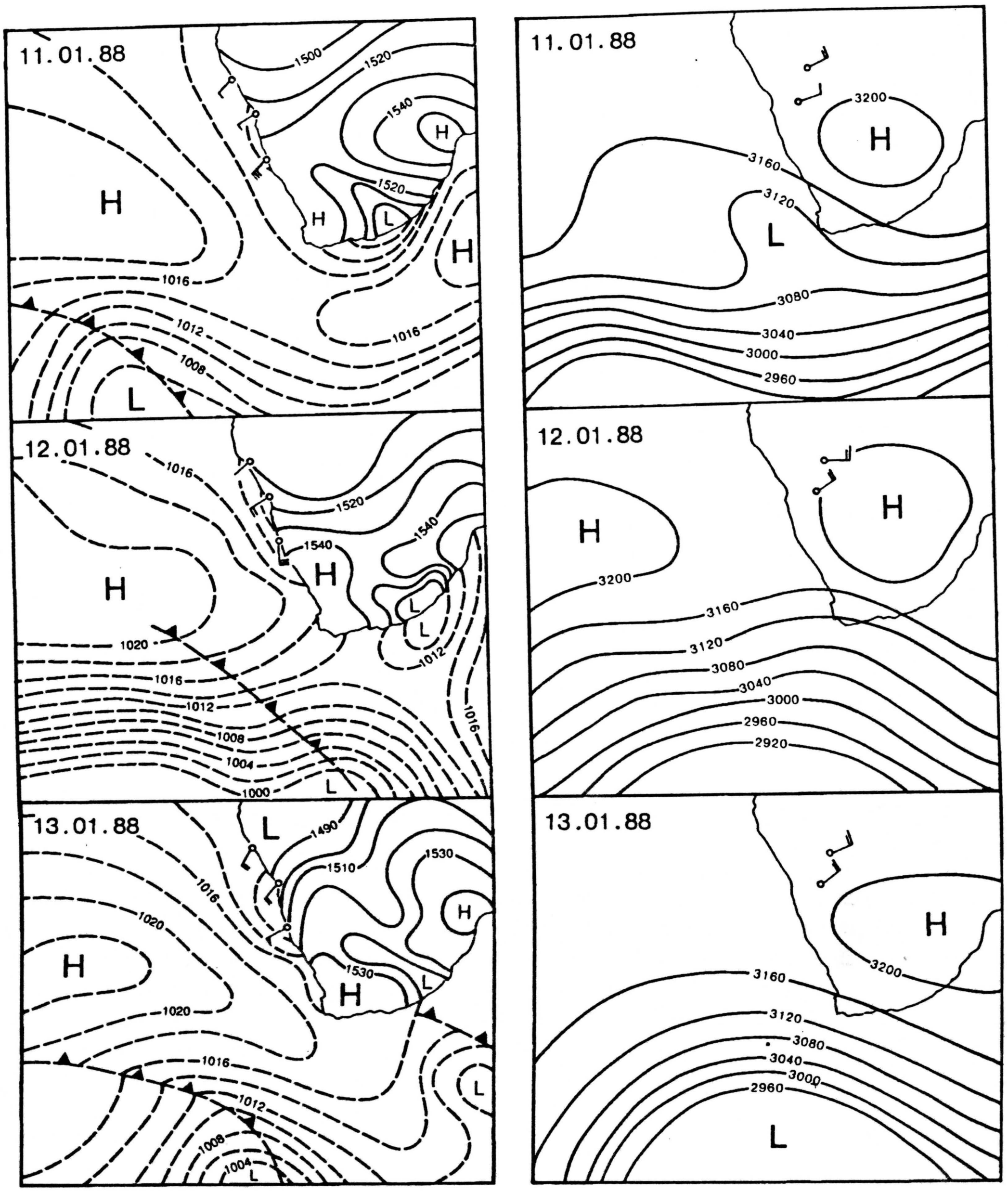

FIGURE 8: Surface (left) and 700 hPa (right) synoptic charts at 14:00 for the period 11-13 January 1988. Pecked lines on the surface charts represent isobars of sea-level pressure; solid lines over the subcontinent are contours of the 850 hPa suface (in gpm). All lines on the 700 hPa charts are contours (gpm). Winds are indicated by flags flying with the wind.

westerly prior to the approach of the system, through north-easterly to easterly with the arrival of the system, to northerly to north-westerly once the low centre has moved south of the observation point (Preston-Whyte 1975, Diab 1986) (Fig. 4). Between 11 and 14 January 1988, however, although the normal diurnal wind pattern was disrupted, the expected coastal low-associated wind changes did not occur. Winds were not markedly north-easterly at Gobabeb on 11 January (Fig. 5), as would be expected with the approach of the coastal low (Fig. 4); the first indication of the disturbance was the strong south-westerly flow on 12 January (Fig. 5). A characteristic of the summer west coast low is that it is initiated by the eastward ridging of an anticyclone to the south of the subcontinent, and the subsequent movement of the coastal low depends on the extent of the high pressure ridging. With the eastward movement of the anticyclone the low moves southward, and has even been known to move northward if the anticyclone strengthens (Diab 1986). The synoptic sequence for 11-13 January 1988 (Fig. 8) shows that the disturbance, which was probably initiated as a coastal low, was prevented from following a path along the coast by the rapid progression of westerly waves further south which did not allow the eastward ridging of the anticyclone.

More detailed synoptic analysis using satellite-derived information (Fig. 9) reveals that the 700 hPa tropical circulation over the area between 11 and 12 January was characterised by the westward migration of a trough axis which crossed the Escarpment around midnight on 11 January (Fig. 9a). This easterly disturbance, developed above the plateau level and not evident on the standard synoptic charts, therefore moved into a position where it could influence the further evolution and behaviour of the coastal disturbance from that time. The low was spatially anchored by the circulation to the south, and the disturbance was then enhanced and prolonged by the interaction between the original low and the easterly wave disturbance to the north. By definition the system was not,

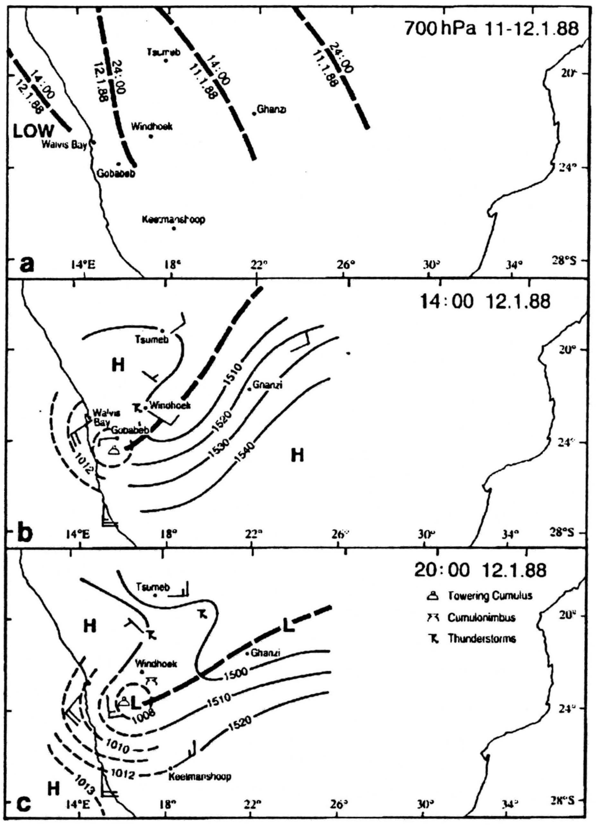

FIGURE 9: Detailed meteorological observations for central Namibia and the west coast, 11-12 January 1988 (data from the South African Weather Bureau). (a) Westward progression of a trough axis (heavy pecked line) at 700 hPa; (b) and (c) sea-level pressure (pecked lines) and 850 hPa contours (solid lines) at 14:00 and 20:00 on 12 January 1988. The heavy pecked line indicates the trough axis, conventional symbols show cloud and thunderstorm activity, and winds are indicated by flags flying with the wind. Sea-level pressure is hPa and geopotential heights are gpm.

therefore, a true coastal low, and the observed disruption of boundary layer conditions during January 1988 can not be ascribed simply to the development and passage of a coastal low. Instead a complex interaction among three aspects of the synoptic circulation, i.e. the initial coastal low, the ridging anticyclone and midlatitude depressions to the south, and the passage of a tropical easterly wave to the north, produced the observed disturbance of the regular diurnal rhythm of summer boundary layer airflows over the central Namib.

CONCLUSION

The central Namib is an ideal area for the study of the inland extent of the influence of synoptic-scale disturbances such as coastal lows. The particular combination of physical and synoptic features in the area is conducive to the formation of coastal lows, and the pronounced pattern of diurnal thermo-topographically induced winds serves as a sensitive background for the identification of disturbances. The synoptic circulation in the area affects the strength of the cyclonic curvature of the developing system, and thus influences the extent and type of disturbance to the boundary layer circulation experienced inland. In the instance documented in this paper the boundary layer disturbance was due to the interaction between a semi-stationary coastal low pressure system, spatially anchored by the subtropical circulation to the south, and an easterly trough above the Escarpment.

In the presence of the disturbance south-westerly winds were stronger and measured further inland than is usually the case in summer. No clear pattern was evident in the disruptive winds, due mainly to the manner in which the disturbance evolved from a coastal depression to a system embedded within an easterly trough. The influence of the disturbance diminished first at Zebra Pan, the station furthest inland, and was most clearly observed at Gobabeb in the central coastal plain. With the unique opportunity presented by the recording of boundary-layer winds over the central Namib during the passage of a synoptic-scale disturbance, it has been possible to show the extent of the disruption of the normal diurnal progression of winds in this region by such a system. It is also clearly evident that the nature of the disruption is dependent upon atmospheric processes beyond the immediate area of the central Namib and on the complex interaction among features of the tropical and subtropical circulations.

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