Global Warming, El Niño, and High-Impact Storms at Extreme Altitude: Historical Trends and Consequences for Mountaineers

G. W. K. Moore
Department of Physics, University of Toronto, Toronto, Ontario, Canada

J. L. Semple
Department of Surgery, University of Toronto, Toronto, Ontario, Canada

G. Hoyland
High Peak, Derbyshire, England

(Manuscript received 25 January 2011, in final form 20 June 2011)

ABSTRACT

The twentieth century was bracketed by two high-profile events on Mount Everest: the 1924 Mallory and Irvine disappearance and the 1996 Into Thin Air storm. During both events, fatalities occurred high on the mountain during deteriorating weather conditions. Although there have been dramatic improvements in knowledge of the mountain and in the technology used on it, it is shown that an unappreciated change that has also occurred, as a result of warming in the region, is an increase in barometric pressure. A rare and unique set of meteorological data collected at various elevations on the mountain during the 1924 British Everest expedition as well as modern datasets are used to compare and contrast conditions during the two storms and the two climbing seasons. It is shown that both storms were associated with weather systems known locally as western disturbances that resulted in summit barometric pressure drops sufficient to have exacerbated altitude-induced hypoxia. It is further shown that the Mallory and Irvine attempt occurred later in the season than typically is the case now and that this was most likely the result of a concurrent El Niño event. Despite the trend of increasing barometric pressure, the pressure drop associated with storms in the region should remain a concern for those who venture to extreme altitudes. The authors therefore argue that success and failure on Everest and other Himalayan peaks requires knowledge of the variability and trends in both the weather and climate.

1. Introduction

The 1924 British Everest expedition represented the culmination of the early attempts to summit Mount Everest (Venables 2003). The expedition arrived at Base Camp, 5029 m MSL, on the north (Tibetan) side of the mountain on 1 May 1924 (Fig. 1). During May, progress was thwarted by the passage of a number of weather systems that brought cold temperatures and snowy conditions to the region (Norton 1925). A period of better weather near the end of May and the beginning of June allowed for two summit attempts. On 4 June, Colonel Edward Norton, supported by Howard Somervell, reached an altitude of 8570 m without the use of supplementary oxygen, an altitude record that was to survive for over 50 years (West 2000). Subsequent to that attempt, George Mallory and Andrew “Sandy” Irvine on 8 June made an attempt using supplementary oxygen. On that afternoon, snow and strong winds engulfed the mountain in a storm that was described at the time as being a “rather severe blizzard” (Odell 1924). Mallory and Irvine never returned, and over the years a number of artifacts from their climb have been recovered, with the most significant being the 1999 discovery of Mallory’s body (Anker and Roberts 1999). Despite these discoveries, much is still not known about their climb, including whether they were successful in their summit attempt.

Seventy-three years later on the evening of 9 May 1996, a large number of climbers were poised to make...
summit attempts, having climbed from the Nepalese Base Camp to the South Col (8000 m MSL) of Mount Everest (Fig. 1). High winds had persisted throughout the day, and the possibility of summiting appeared to be low (Krakauer 1999). The winds died down during the evening, and the conventional wisdom was that they would remain calm for a period of time. As a result, the decision was made to attempt to summit (Krakauer 1999). During the afternoon of 10 May, an intense storm, with wind speeds estimated to be in excess of 30 m s$^{-1}$, heavy snowfall, and falling temperatures, engulfed Mount Everest (Krakauer 1999). More than 20 climbers were trapped on its exposed upper slopes. The high winds and continuing harsh weather hampered a number of attempts to rescue the trapped climbers. Eight of the climbers perished during this storm, which is the highest number to die during a single event on Mount Everest.

The 1924 expedition’s data collection activities represented some of the earliest meteorological measurements collected in this remote region that even today has a dearth of observations. The focus of interest was on simultaneous temperature observations at elevations up to 7000 m MSL to calculate the environmental lapse rate, the rate at which temperature decreases with height (Somervell 1926). Daily barometric pressure data were also collected at Base Camp (Somervell 1926). It is difficult to ascertain the uncertainty in these historical measurements, but the care with which the data were collected, along with evidence of calibration of the instruments before and after the expedition, suggests that they were indeed representative of actual conditions (Moore et al. 2010).

The temperature data indicated that the average environmental lapse rate at local noon was approximately 8°C km$^{-1}$, indicating that the atmosphere was stably stratified (Whipple 1926; Holton 2004). The barometric pressure data from the 1924 expedition were published as a table in 1926 (Somervell 1926) but were not until recently analyzed to provide information on the storm that occurred during the Mallory and Irvine climb (Moore et al. 2010).
et al. 2010). This analysis showed that there was an 18-hPa drop in barometric pressure at Base Camp during the Mallory and Irvine summit attempt. The large magnitude of the drop suggested that the weather during their attempt was more severe than was originally assumed and that it most likely contributed to their deaths (Moore et al. 2010).

Using a contemporary manual sea level pressure analysis from the Indian Meteorological Department, Moore et al. (2010) showed that during the 1924 attempt there was a region of low pressure to the west of Mount Everest with southerly flow into the region from the Bay of Bengal. They furthermore argued that the coupling of this surface feature with the upper-level trough responsible for the drop in barometric pressure observed at Base Camp is similar to that associated with so-called western disturbances that are responsible for much of the cool-season, that is, premonsoon, severe weather in the Himalaya (Dimri 2004; Lang and Barros 2004).

Moore and Semple (2006) used the 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40; Uppala et al. 2005) to investigate the meteorological conditions during the 1996 storm. They found that the storm was associated with a drop in summit barometric pressure of approximately 6 hPa and that it was the result of the juxtaposition of two jet streaks—regions of high winds embedded in the subtropical jet stream—that led to an outbreak of organized convective activity that engulfed the mountain.

A surface temperature reconstruction based on glacier-length data indicated that the high mountainous regions of Asia have warmed by approximately 0.6°C since the middle of the nineteenth century (Oerlemans 2005). There is evidence of a recent acceleration in this warming over the southern Tibetan Plateau, with annual mean surface air temperatures increasing by approximately 0.2°C (10 yr)−1 since the 1960s (Lin et al. 2010). Over the past 30 years, this warming has resulted in a 100–300-m rise in the permafrost height in the Mount Everest region (Fukui et al. 2007). The annual mean surface pressure at Lhasa, Tibet (3658 m MSL), has been increasing at a rate of approximately 0.3 hPa (10 yr)−1 since the 1950s—an increase that has been proposed to be associated with regional warming (Toumi et al. 1999). Moore and Semple (2009) argued that since 1948 this warming has also increased the barometric pressure at Everest’s summit. For much of the year, the statistically significant trend in summit barometric pressure is on the order of 0.2–0.3 hPa (10 yr)−1. Moore and Semple (2009) further argue that this increase is of a magnitude to be of physiological significance to climbers attempting to summit.

Of the approximately 3000 successful ascents of Mount Everest, 83% have occurred in May and only 4% in June, with all of the latter taking place during the first 5 days of that month (Hawley and Salisbury 2007). This is because the onset of the Indian summer monsoon typically brings heavy snowfall to the region in early June that makes it dangerous to be on the mountain (Venables 2003). Indeed, the 1922 Mount Everest expedition ended abruptly when heavy snowfall, attributed to the monsoon, that fell on 3 and 4 June caused an avalanche that resulted in the deaths of seven porters (Bruce 1923).

It is therefore clear that the 1924 summit attempts, on 4 and 8 June, occurred later than is typically the case at present. As we will argue in this paper, one intriguing possibility that may explain the lateness of the 1924 summit attempts is related to a concurrent El Niño event. El Niño events are associated with anomalous warming of the sea surface in the eastern tropical Pacific Ocean (Trenberth 1997) and typically lead to an increase in Eurasian snowfall, including in Tibet (Dong and Valdes 1998), and a reduction in the intensity of the Indian summer monsoon (Webster and Yang 1992). It has been recently proposed that there are two distinct types of both warm El Niño and cold La Niña events. In addition to conventional events that have sea surface temperature extrema in the eastern tropical Pacific, there are also events for which the extrema are situated in the central tropical Pacific close to the international date line (Trenberth and Stepaniak 2001; Ashok and Yamagata 2009). El Niño events associated with a reduction in the intensity of the Indian summer monsoon are typically of the newly identified central Pacific type (Kumar et al. 2006).

In this paper, we use the meteorological data collected during the 1924 expedition as well as modern datasets to compare and contrast conditions during the two storms and the two climbing seasons. We will show that, although regional warming has resulted in an increase in surface pressure over Tibet, the pressure drops as well as the pressure minimum during both the 1924 and 1996 storms were of similar magnitudes. In addition, we will show that the sea surface temperature anomaly during the 1924 El Niño event was centered in the central tropical Pacific and therefore was of the type associated with a reduction in the intensity of the Indian summer monsoon. This result supports the hypothesis that the 1924 El Niño contributed to a delay in the onset of the monsoon, therefore allowing the lateness of the summit attempts during the 1924 expedition.

2. Data and methods

During the 1924 expedition, surface temperature measurements were made at three times per day (0830, 1200, and 1600 local time) at elevations from Base Camp.
to Camp IV at 7010 m MSL (Somervell 1926). In addition, barometric pressure measurements were made each day at 0830 local time at Base Camp (Somervell 1926). For this paper, we will use the 1924 observed average environmental lapse rate of 8°C km⁻¹ to integrate the 1924 barometric pressure and temperature observations from Base Camp to the summit of Mount Everest at 8848 m MSL. Appendix A describes the details of the method used.

The 1924 results will be compared with time series of summit temperature and barometric pressure for the 1996 climbing season as extracted from the National Centers for Environmental Prediction (NCEP) reanalysis (Kalnay et al. 1996). Reanalyses make use of modern numerical weather prediction systems to assimilate historical observations into consistent and homogeneous datasets suitable for diagnostic and model-validation studies (Kalnay et al. 1996; Uppala et al. 2005). Comparisons with observations at 5800 and 8000 m in the vicinity of Mount Everest indicate that the NCEP reanalysis is able to capture the synoptic-scale and seasonal pressure and temperature variability in the region (Moore and Semple 2004; Xie et al. 2009). The NCEP reanalysis data are available for four times per day, and data from 1200 UTC, the time closest to that at which the 1924 observations were made, were used for this comparison. The 1924 data are available from 1 May to 13 June, and this period was selected for the comparison. Climatological measures of the summit temperature and barometric pressure were also calculated from the NCEP reanalysis for the period 1948–2009.

We also use the sea level and surface pressure fields from the Twentieth-Century Reanalysis Project (20CR) dataset (Compo et al. 2011) to assess conditions during the 1924 and 1996 storms and climbing seasons. The 20CR assimilates only surface pressure observations and uses an ensemble of 56 short-term parallel integrations of the atmospheric component of the NCEP Climate Forecast System model to achieve the first three-dimensional representation of the state of the troposphere for 1871 onward. The 20CR includes an estimate of the state of the atmosphere, provided by the mean over the ensemble members, as well as a measure of the uncertainty in this estimate, provided by the standard deviation among the ensemble members.

The trend in surface pressure from the 20CR was also calculated to assess the long-term changes over the Tibetan Plateau associated with global warming. Although there is some controversy about the use of reanalyses to estimate climate trends (Bengtsson et al. 2004; Thorne and Vose 2010; Dee et al. 2011), there is nevertheless evidence that, including the 20CR, contain trend information that is consistent with other datasets (Simmons et al. 2004, 2010; Bengtsson and Hodges 2011; Compo et al. 2011).

Geophysical time series tend to be temporally autocorrelated, resulting in so-called red-noise behavior that leads to a reduction in the degrees of freedom (Ghil et al. 2002). To take this into account, the statistical significance of the trend was estimated with a resampling technique that randomizes the phase of the components of the Fourier decomposition of surface pressure time series to generate 1000 synthetic time series that preserve the spectral characteristics of the original time series (Rudnick and Davis 2003). The distribution of the trends from these synthetic time series was then used to estimate the statistical significance of the 20CR trend at each grid point (Moore and Semple 2009).

We also use a reconstruction of the sea surface temperature field (Kaplan et al. 1998) that includes all available surface maritime measurements to diagnose the state of the El Niño–Southern Oscillation (ENSO) during 1924 and 1996.

3. Results

The 1924 and 1996 summit temperature time series are shown in Fig. 2. Both time series show an increasing trend over the period of interest that is consistent with the warming in the region that occurs during the spring (Moore and Semple 2004; Xie et al. 2009). During 1924, the summit temperature was, on average, below the modern climatological mean, with the largest deviation occurring during the first two weeks of May when it was −3.6°C. During 1996, the summit temperatures were closer to but still, on average, lower than the climatological values, with the average deviation on the order of −1°C.

The 1924 and 1996 summit barometric pressure time series are shown in Fig. 3. Both show a tendency for the barometric pressure to increase through May and June that is a result of the aforementioned springtime warming that results in a thickening of the atmosphere (Moore and Semple 2004; Xie et al. 2009). During 1924, the summit barometric pressure was, on average, lower than the modern climatological mean, with the magnitude of the difference decreasing through May and June. For example, the difference was −6 hPa during May and decreased to −3 hPa during the first two weeks of June. The 1924 storm occurred during a period in which the summit barometric pressure fell from 341 hPa on 6 June to 331 hPa on 9 June, a drop of 10 hPa. During 1996, the summit barometric pressure was close to the climatological mean, with an average difference of −0.3 hPa. The 1996 storm also occurred during a period of falling pressure, from 337 hPa on 7 May to 331 hPa on 12 May, a drop of 6 hPa.
In Fig. 4, we show the sea level pressure field during the 1924 and 1996 storms as expressed in the 20CR. The surface circulation during the 1996 storm was similar to that during the 1924 storm, with a low pressure center to the west of Mount Everest and southerly flow into the region from the Bay of Bengal. The low pressure system was deeper in 1924 than in 1996, however. Given the similarity between the two sea level pressure fields, it is therefore likely that the 1996 storm was also a western disturbance.

Figure 5 shows the monthly mean surface pressure anomalies from the 20CR for May and June during 1924 and 1996. During May of 1924 (Fig. 5a), there was a well-defined dipolar anomaly in surface pressure, with higher values over northwest India and lower values over Tibet. During May of 1996 (Fig. 5b), the magnitudes of the anomalies were smaller and did not exhibit the same degree of organization that was seen during May of 1924. The anomaly during June of 1924 (Fig. 5c) had higher surface pressures over and north of the Tibetan Plateau, with lower surface pressures over the Bay of Bengal. As was the case during May of 1996, the anomaly field during June of 1996 (Fig. 5d) lacked spatial coherence.

The trend in the annual mean surface pressure field from the 20CR over the period 1900–2008 is shown in Fig. 6. Over the twentieth century there has been a redistribution of atmospheric mass in the region, with a statistically significant increase in surface pressure over the Tibetan Plateau and a statistically significant decrease in the surrounding regions. Over the plateau, the trend in surface pressure was on the order of 0.05–0.1 hPa (10 yr)\(^{-1}\) over this period.

Figure 7 shows the sea surface temperature anomaly for the 3-month period from January to March during both 1924 and 1996. During 1924 (Fig. 7a), there was a warm anomaly in the central-eastern tropical Pacific,
with anomalously cold sea surface temperatures along the South American coast. In contrast, 1996 was associated with cold sea surface temperature anomalies across the entire eastern tropical Pacific from the coast of South America to the international date line. The Niño-3.4 index, defined as the anomaly in the sea surface temperature in a region bounded by 5°S–5°N and 120°–170°W (Trenberth 1997), was 0.72° and −0.68°C during these 3-month periods in 1924 and 1996, respectively. According to the most commonly accepted definition of ENSO events (Larkin and Harrison 2005), 1924 meets the criteria for a warm El Niño event and 1996 meets the criteria for a cold La Niña event. The identification of 1924 as an El Niño event is consistent with independent assessments that use different multivariate criteria (Quinn et al. 1978; Kiladis and van Loon 1988). The dipole structure of the anomaly during 1924, with a warm anomaly in the central-eastern Pacific and a cold anomaly along the coast of South America, is that associated with the newly identified central Pacific class of El Niño events (Trenberth and Stepaniak 2001; Ashok and Yamagata 2009) that are associated with a reduction in the intensity of the monsoon (Kumar et al. 2006).

4. Discussion

As we have shown, both the 1924 and 1996 storms were associated with summit barometric pressure drops that were most likely associated with synoptic-scale low pressure systems known locally as western disturbances (Figs. 3, 4). The pressure drop was larger and occurred more quickly for the 1924 storm, suggesting that it may have been more intense than the 1996 storm. This conclusion is also consistent with the sea level pressure fields during the two storms (Fig. 4).

The 9 June 1924 sea level pressure field from the 20CR (Fig. 4a) is similar to the contemporary manual analysis from the Indian Meteorological Department shown in
Moore et al. (2010), except for a slight change in the orientation of the region of low pressure as well as an approximately 2-hPa difference in the central pressure. The NCEP reanalysis sea level pressure field during the 1996 storm was similar to that of the 20CR but with the depth of the low pressure system being approximately 2 hPa shallower. These differences are consistent with the uncertainty estimates of the 20CR.

The minimum summit barometric pressure was approximately 331 hPa during both the 1924 and 1996 storms (Fig. 3). As described in more detail in appendix B, this is close to the human limit for tolerance to hypoxia. When at extreme altitude (above 7000 m), the ability to acclimatize is minimized and changes in barometric pressure on the order of 4 hPa can have a significant effect on a climber’s physiology (West 1983). It is clear that both storms were associated with summit barometric pressures and drops that were of a magnitude to drive individuals into a profoundly hypoxic state. These results are consistent with a recent analysis of mortality on Mount Everest, which indicated that deaths attributable to the weather were typically associated with a drop in summit barometric pressure (Firth et al. 2008).

Western disturbances typically have their largest amplitude in the midtroposphere at approximately the 500-hPa level (Lang and Barros 2004). This is consistent

FIG. 4. The sea level pressure field (hPa) from the 20CR for (a) 0600 UTC 9 Jun 1924 and (b) 0600 UTC 11 May 1996. The extent of the Tibetan Plateau is indicated by the thick blue 3000-m isocontour. The location of Mount Everest is indicated by the asterisk.
with the results that we have obtained in which the 1924 pressure drop at Base Camp, which is located at approximately 500 hPa, was larger than that at the summit of Mount Everest. This result also follows from the expressions for the summit barometric pressure in terms of the pressure and temperature observed at Base Camp [appendix A Eqs. (A4) and (A5)]. By assuming, for simplicity, an isothermal atmosphere, one obtains the result

$$\Delta P_{\text{isothermal}}(z) \approx \Delta P_0 e^{-\frac{(z-z_0)}{H}}. \quad (1)$$

Therefore the changes in barometric pressure at the summit should be \(\exp[-(z - z_0)/H]\), or \(\approx 0.6\) as large as those at Base Camp. This is close to the observed ratio. A similar result is obtained if one assumes a constant stably stratified lapse rate.

With regard to mean conditions, it is clear from the results presented that May of 1924 had anomalously low summit temperatures and barometric pressures whereas June of 1924 was closer to but still below modern mean values (Figs. 2, 3). In this regard it is interesting to note that the 1924 expedition report states that May was characterized by cold temperatures and stormy weather that according to Darjeeling tea planters were unique in the preceding 20–50 years (Norton 1925). The monthly mean surface pressure anomaly during May of 1924 (Fig. 5a) is consistent with these observations in that surface pressures over Tibet were lower by as much as 2 hPa. The year 1924 was also anomalous in that the summit attempts occurred later in June than is generally the case now for springtime ascents. We have shown that during June of 1924 surface pressures over the Tibetan Plateau were as much as 1 hPa higher than usual (Fig. 5c). This anomalously high pressure was most likely associated with a delay in the onset of the Indian summer monsoon that resulted from the concurrent El Niño event and contributed to the late ascents. In contrast, mean conditions during May and June of 1996 (Figs. 5b,d) did not exhibit the same degree of spatial coherence, suggesting that there was not the same degree of large-scale influence as was the case during 1924.

Figure 6 indicates that over the period from 1900 to 2008 there has been a tendency toward an increase in surface pressure of approximately 0.1 hPa (10 yr\(^{-1}\)) over much of the Tibetan Plateau. This result is consistent with the trend in surface pressure at Lhasa (Toumi et al. 1999).
and in northwest India (not shown). Assuming an isothermal atmosphere [appendix A Eq. (A5)], this change in surface pressure $P_s$ at a constant height $z$ can be related to a change in surface temperature $T_s$ through

$$\Delta P_s \approx \frac{P_s}{T_s} \frac{z}{H} \Delta T_s,$$  \hspace{1cm} (2)

with the result being that the impact of a surface warming on surface pressure increases with increasing height (Toumi et al. 1999), suggesting that surface pressure variations over the Tibetan Plateau are an excellent proxy for surface temperature variations in the region. In this regard, the 20CR, which assimilates only surface pressure data, offers the possibility of generating

![Figure 6. The trend [hPa (10 yr)$^{-1}$] in annual mean surface pressure from the 20CR for the period from 1900 to 2008. The shading represents regions where the trend is statistically significant at the 5% level. The extent of the Tibetan Plateau is indicated by the thick blue 3000-m isocontour. The location of Mount Everest is indicated by the asterisk.](image)

![Figure 7. The anomaly in the sea surface temperature field ($^\circ$C) from the Kaplan et al. (1998) dataset for January–March (a) 1924 and (b) 1996.](image)
information on regional surface warming over its 134-yr length (1871–2008). Assuming a surface temperature trend over the plateau during the twentieth century that is on the order of 0.1°–0.2°C (Lin et al. 2010), this implies a surface pressure trend on the order of 0.1–0.2 hPa (10 yr)^{-1}, which is a result that is consistent with the trend in Fig. 6.

These trend results and those of Moore and Semple (2009) imply that the mean summit barometric pressure in 1924 should have been on the order of 2 hPa below current values. The summit pressure during May was as much as 6 hPa below the modern climatological mean, and that for June was 3 hPa below it. This suggests that the discrepancy in June can be attributed to the long-term warming in the region, whereas there must have been other factors that contributed to conditions in May.

As we have shown, 1924 was an El Niño year (Fig. 7a). El Niño events typically lead to an increase in Tibetan snowfall (Dong and Valdes 1998), and, as we have shown, May of 1924 had anomalously low surface pressures over the plateau (Fig. 5a). Furthermore, winter and spring Tibetan snow cover is usually anticorrelated with the intensity of the following Indian summer monsoon as a result of a reduction in the ocean–land temperature contrast that drives the monsoon (Blanford 1884; Dong and Valdes 1998). In addition, the changes in atmospheric circulation associated with El Niño events result in summertime subsidence over India that also typically reduces intensity of the Indian summer monsoon (Kumar et al. 2006). In this regard, we have shown that surface pressures during June of 1924 were anomalously high (Fig. 5c). El Niño events that are most strongly correlated with a reduction in the monsoon typically have their largest sea surface temperature anomalies in the central-eastern Pacific (Kumar et al. 2006), and, as shown in Fig. 7, this was the case during 1924.

It is interesting to note that 1996 was a La Niña year and so, assuming a linear response, one might expect opposite impacts during this climbing season. Figure 5 indicated that no coherent patterns in the surface pressure anomaly field were present during May and June 1996, however. There are two possible explanations that may explain this discrepancy. There is evidence of a non-linearity in the response to El Niño and La Niña events (Hoerling et al. 1997). In addition, the 1996 La Niña event was not of the central Pacific type (Fig. 7b) that has been identified to have the largest impact on the Indian subcontinent.

5. Conclusions

We have investigated the meteorological conditions during the 1924 and 1996 storms and climbing seasons with a view toward understanding their similarities and differences as well as characterizing the impact that global warming and ENSO have on the meteorological behavior and climate of the Mount Everest region.

Both the 1924 and 1996 storms were associated with large drops in barometric pressure and appear to be manifestations of synoptic-scale low pressure systems known locally as western disturbances that are responsible for much of the cool-season severe weather in the Himalaya (Dimri 2004; Lang and Barros 2004). The role of western disturbances as a source of convective weather in northern India and Nepal has been known since the early period of synoptic meteorology in India (Blanford 1884). The observation that both of these high-profile storms had a common synoptic-scale signature suggests that forecasts as to development of western disturbances would be of benefit to climbers in the high Himalaya as predictors of the possible development of severe weather.

We have presented evidence that the high elevation of Mount Everest in particular and the Tibetan Plateau in general is such that surface barometric pressure trends are excellent indicators of surface warming. These trends act to increase the barometric pressure and are of a magnitude to be of physiological significance (Moore and Semple 2009). Nevertheless the pressure drop and ensuing bad weather associated with modern-day storms should remain of concern for the increasing numbers of people who venture to extreme altitudes.

The spring climbing season on Mount Everest is usually restricted to April and May and is typically terminated by the arrival of the Indian summer monsoon near the end of May. The summit attempts in 1924 occurred later than is typically the case today, and we have presented evidence that a delay in the onset of the monsoon arising from a concurrent central Pacific El Niño was responsible. It follows that during such years it may be possible to extend the climbing season. Logistical constraints such as the expiration of permits typically do not allow for this degree of flexibility, however.

The identification of 1924 as a central Pacific El Niño year is also of interest because it has been recently suggested that no such events occurred prior to the 1960s and that their recent occurrence is related to global warming (Ashok and Yamagata 2009; Yeh et al. 2009). Our identification of an early-twentieth-century central Pacific El Niño event suggests a reevaluation of this hypothesis.

Our results imply that a complete explanation of past success and failure on Mount Everest and, by implication, other high Himalayan mountains, as well as predictions of future success, clearly requires knowledge not only of the nature and intensity of storms that engulf
the mountain but also of the impacts of climate trends and variability. It follows that climbers who attempt to summit high Himalayan mountains should take into account all available information on weather and climate conditions in the region.

Acknowledgments. GWKM was supported by the Natural Sciences and Engineering Research Council of Canada. The NCEP reanalysis, 20CR, and Kaplan et al. sea surface temperature data used in this paper were provided by the NOAA/OAR/ESRL PSD, which is located in Boulder, Colorado. We thank the editor and reviewers for their comments.

APPENDIX A

Extrapolation of the 1924 Base Camp Observations to the Summit

Under the assumptions that the atmosphere is in hydrostatic balance and is an ideal gas, the variation of barometric pressure $P$ with height $z$ may be written as

$$\frac{\partial P}{\partial z} = -\frac{P}{RT}g, \quad (A1)$$

where $T$ is the atmospheric temperature, $R$ is the gas constant for dry air, and $g$ is the acceleration resulting from gravity (Holton 2004). The environmental lapse rate $\Gamma$ is defined as

$$\Gamma = -\frac{\partial T}{\partial z}. \quad (A2)$$

If one assumes that $\Gamma$ is constant, then Eq. (1) may be integrated with boundary conditions

$$P = P_o \{ \begin{array}{c} 
T = T_o 
\end{array} \} \text{ at } z = z_o \quad (A3)$$

to obtain the following expressions for the variation of barometric pressure and temperature with height:

$$T(z) = T_o - \Gamma(z - z_o) \quad \text{ and}$$

$$P(z) = P_o [1 - \Gamma(z - z_o)/T_o]^{g/(TR)}. \quad (A4)$$

These expressions along with the observed value of $\Gamma$ were used to integrate the observed temperature $T_o$ and barometric pressure $P_o$ time series obtained at Base Camp ($z_o = 5029 \text{ m}$) to the summit of Mount Everest ($z_s = 8848 \text{ m}$).

As a test of the uncertainty in using the observed environmental lapse rate $\Gamma$ of $8^\circ C \text{ km}^{-1}$, the integrations were repeated while assuming a dry adiabatic lapse rate $\Gamma_a$ of $9.8^\circ C \text{ km}^{-1}$ (Holton 2004) as well as assuming an environmental lapse rate of $5^\circ C \text{ km}^{-1}$. As a further check, under the assumption of an isothermal atmosphere, Eq. (1) may be integrated with boundary conditions provided by Eq. (A1) to obtain

$$T_{\text{isothermal}}(z) = T_o \quad \text{ and}$$

$$P_{\text{isothermal}}(z) = P_o e^{-(z-z_o)H}, \quad (A5)$$

where $H = RT_o/g$.

Figure A1 shows the sensitivity of the summit barometric pressure to assumptions regarding the thermal state of the atmosphere for a fixed barometric pressure at Base Camp. As expected, the assumption of an isothermal atmosphere results in higher summit barometric

![Fig. A1. The dependence of summit pressure as a function of Base Camp temperature under different assumptions regarding the thermal structure of the atmosphere. In all cases, the barometric pressure at Base Camp was assumed to be the mean during the 1924 expedition of 549 hPa.](image-url)
pressures as the warmer layer temperature results in a thicker atmosphere. As the lapse rate increases from 5° to 9.8°C km⁻¹ there is a decrease in summit barometric pressure as the increasingly colder mean temperature in the layer results in a thinning of the atmosphere. For the same reason, a decrease in temperature at Base Camp results in a decrease in summit barometric pressure in all cases. For the mean Base Camp temperature, 275.4°C results in a decrease in summit barometric pressure in all cases. For the mean Base Camp temperature, 275.4°C results in a decrease in summit barometric pressure in all cases.

The 1924 observations only reported on the environmental lapse rate to a height of approximately 7000 m MSL. There is unfortunately no information on the lapse rate above this height available (either currently or from the 1924 dataset). The simplest assumption is that the lapse rate is the same above this height as it is below. As a further test, the lapse rate above 7000 m was varied to see the impact on the summit barometric pressure. As shown in Table A1, the impacts are small for the realistic cases in which one assumes a constant lapse rate but are much larger for the case in which one assumes an isothermal atmosphere above 7000 m.

APPENDIX B

Human Limits to Hypoxia

A useful measure of the body’s physiological response to the low levels of oxygen available at high altitude is the maximal oxygen uptake \( V_{O_{2 \max}} \) (West 2000). Fit individuals at sea level have a \( V_{O_{2 \max}} \) of approximately 4500 ml min⁻¹; at a barometric pressure of 333 hPa, it is reduced to approximately 1000 ml min⁻¹ (West et al. 1983). On the basis of exercise studies during the U.S. medical research expedition to Everest, \( V_{O_{2 \max}} \) varies with barometric pressure near the summit of Mount Everest at an approximate rate of 10 ml min⁻¹ hPa⁻¹ (West 1983). Therefore, a drop in summit barometric pressure of 6–10 hPa, as occurred during the 1924 and 1996 storms, would result in a 5%–10% drop in \( V_{O_{2 \max}} \). Given the delicate balance that exists with respect to hypoxic stress near the summit, a drop of this magnitude would make it unlikely that the mountain could be climbed without the use of supplementary oxygen (West 1983).

The impact of supplementary oxygen on Mount Everest has been controversial since the earliest attempts to climb Mount Everest in the 1920s (Windsor et al. 2008). At the summit of Mount Everest, supplementary oxygen typically increases \( P_{O_{2}} \) by up to 13 hPa, which, because of the steepness of the \( P_{O_{2}} \)-to-\( V_{O_{2 \max}} \) curve results in \( V_{O_{2 \max}} \) being as large as 1600 ml min⁻¹ (West et al. 1983; Windsor et al. 2007). This is equivalent to \( P_{O_{2}} \) and \( V_{O_{2 \max}} \) for unassisted breathing at a barometric pressure of approximately 380 hPa or a height of approximately 7800 m (West et al. 1983). At heights above approximately 7000 m, there is significant hypoxic stress (West 2000); as a result, even with supplementary oxygen climbers are still prone to hypoxia. In the cases of both the 1924 and 1996 events, however, it is clear that reserves of supplementary oxygen were exhausted early in the storms (Hemmleb et al. 1999; Krakauer 1999), thereby exposing individuals to the full effect of the lower barometric pressure.

REFERENCES


| TABLE A1. Root-mean-square error (RMSE; hPa) between the summit barometric pressure calculated using the observed environmental lapse rate of 8°C km⁻¹ from Base Camp to the summit and various other scenarios with respect to the environmental lapse rate. |
|---|---|---|
| 8°C km⁻¹ to 7000 m and then isothermal atmosphere | 8°C km⁻¹ to 7000 m and then 5°C km⁻¹ | 8°C km⁻¹ to 7000 m and then 9.8°C km⁻¹ |
| RMSE | 2.40 hPa | 0.92 hPa | 0.62 hPa |