# 2 The Mediterranean environment

## *2.1 Introduction*

Qualitatively, the Mediterranean climate is typified by long warm and dry, often hot and sunny summers, with occasional showers and thunderstorms (Wallen, 1970). The winter months are cool and variable with periods of dry and cloudy days interspersed by sunny periods or rain. Characteristic of this regime are intense rainy seasons that occur in spring or autumn dependent on location. Gales occur in winter, in some places quite frequently, and along coastal areas and the Spanish valleys fog may occasionally appear (Rudloff, 1981).

A more quantitative definition, and an example of 'climate classification' is given in Table 2.1, based on the work of Köppen (1931). Köppen's work, a classification of all world climates based upon temperature and precipitation and delimited by spatial variations in flora, was the first to formally introduce the idea of a strictly 'Mediterranean climate' ('Cs', see Table 2.1), consistent with the above qualitative description. It is an example of one of five types (A-E, tropical to polar) of climate principally defined by temperature, and further subdivided according to precipitation and seasonality. The flora that delimits the Mediterranean type is generally hardy, able to cope with low levels of moisture and to make good use of the high levels of sunlight available. Woody scrub, olive groves, citrus plantations and small regions of deciduous forest are all apparent, illustrating the range of landuse found within the basin (Mairota *et al.*, 1998; Bolle, 2003).

Rudloff's 'World Climates' (1981), utilising an adapted version of the Köppen system (the Köppen/Trewartha classification), defined the following areas as possessing a 'Cs' climate; the Mediterranean coast of France and the French hinterland, Corsica, Sicily, Sardinia, the greater part of Spain except northern Spain and parts of eastern Spain, the Balearic Islands, Gibraltar, southern Portugal, the Madeira Islands, southern Italy, Malta, Greece, most parts of Turkey, Cyprus, northern and western Syria, The

Lebanon, northern Israel, and northern Algeria (Fig. 2.1). Examples of the Mediterranean climate that exist outside of the basin, and that satisfy this and the original Köppen definition, can be found in central and western California, the south of the Western Cape province in South Africa and along the western-facing coasts of south western Australia and Central Chile (Trewartha, 1961).

In this study, the European or near-European countries are of particular climatic and socio-economic interest, and the study region is defined as the Northern Mediterranean basin (enclosed by 30-45˚N, 10˚W-35˚E) as used by Palutikof *et al.* (1996). This area includes the majority of countries within the Mediterranean that display the Cs climate type[1]. Also included are smaller regions of 'Bs', and 'Cf' climate (see Table 2.1). On the southern fringes of the study region (i.e., Algeria), desert 'Bw' conditions become more prevalent, while the 'Cf' type climate typifies the rest of Europe to the North.

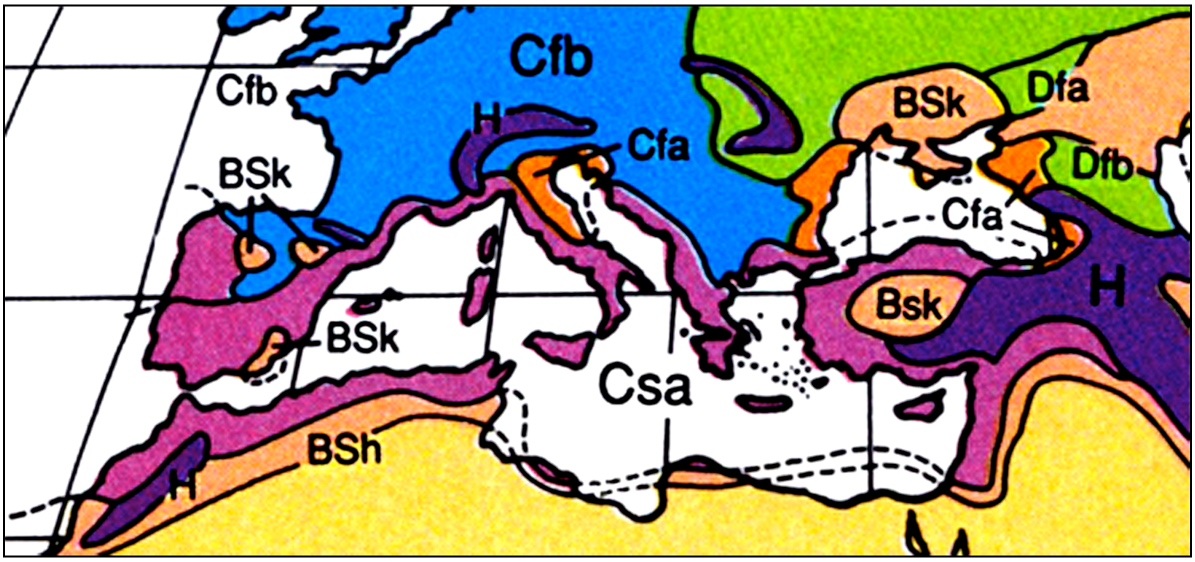


**Figure 2.1: Mediterranean Köppen Climate Classification (from Ahrens, 1991).**

This chapter describes the region of the Mediterranean basin as defined above, and more specifically its climate characteristics. Section 2.2 gives a general description of the Mediterranean climate (Section 2.2.1), and the geographic (Section 2.2.2) and atmospheric (Section 2.2.3) factors that influence it. Section 2.3 provides a more

[1] Socio-economic rationale for the study region can be found in Chapter 5

specific and technical discussion of local atmospheric fields (Section 2.3.1), features (Section 2.3.2), and circulation (Section 2.3.3) and how they are affected by larger scale factors (Section 2.3.4). The role of extreme climate as part of the Mediterranean regime is then dealt with in Section 2.4. Consideration is given to: the difficulties in defining extremes across the region (Section 2.4.1); detecting and studying extremes (Section 2.4.2); how they may be linked to circulatory forcing (Section 2.4.3); and their trends over the last half century (Section 2.4.4). In Section 2.5 the most relevant points to arise from this literature review are summarised.

## *2.2 The Mediterranean in context*

### 2.2.1 The climate

A large inter-seasonal range (between long, dry summers and mild, wet winters) typifies the Mediterranean climate as described in Section 2.1. However, this description lacks detail regarding the features that make Balkan and eastern Mediterranean (Turkey, Cyprus, Crete) weather different from that experienced by either the central Mediterranean (Italy, Sicily, Corsica, Sardinia and the countries bordering the northern Adriatic), the western Mediterranean (the Iberian peninsula, southern France, and the Balearics), or the southern Mediterranean (Algeria, Morocco, Tunisia and Malta) (Wallen, 1970; Rudloff, 1981; Goosens, 1986; Maheras *et al.*, 1999a; Maheras *et al.*, 1999b; Maheras and Kutiel, 1999), regions which fall roughly between 18-30°E, 4-18°E, -10-4°E and below 36°N respectively. The climate of one region is largely different from that of another, although the regions detailed above share some influences (as detailed in Section 2.2.2 and below).

Across the basin there are a number of consistent patterns (Trewartha, 1962):

- Temperatures decrease to the north, with some mitigation by the Mediterranean Sea occurring along coastlines in the winter
- Rainfall decreases southwards in summer. In winter, region-to-region contrasts are less dependent on latitude.

- In the east, minimum temperatures are generally lower than in the west. Invasions of northerly air, which influence Mediterranean minimum temperatures, are colder in the Balkans than in Italy or France.
- Periods of dryness tend to increase west to east as conditions become more continental and less exposed to the Atlantic.
- The orientation of coasts has a large effect on rainfall due to prevailing winds. In both Iberia and Greece, for example, western coasts have double the amount of precipitation found on eastern coasts.

Temperature and precipitation both display high correlations with latitude (negative and positive respectively). A positive relationship between precipitation and longitude also exists in all seasons, although it is statistically significant only in spring and autumn (Trewartha, 1961). In general, the effects listed above are more zonally (north-south) coherent in summer, and more meridionally (east-west) so in winter (see Fig 2.2) (Barry and Chorley, 1998). This reflects the seasonally predominant direction of flow over the region. The annually dominant temperature and precipitation distribution can also be roughly divided into western, central, and eastern groups as above (e.g. Goosens, 1986; Maheras *et al.*, 1999a). In terms of annual total rainfall, Iberia (Spain and Portugal) averages 757 mm (1958-2000 average of 34 stations), a figure that decreases eastwards, with the exception of the western Balkan coast and far-eastern Italy, to a value of 620 mm for Greece (1958-2000 average of 15 stations) (Goosens, 1985). For temperature, a similar west-east polarity is evident (Maheras *et al.*, 1999b), as dictated by changing circulation. A large proportion of annual temperature variance (34.7% of total non-seasonal variance) can be explained by an anti-phase relationship between the eastern and western basin (Corte-Real *et al.*, 1995). During summer, the majority of temperature and precipitation contrasts are produced by localised, stable, thermally-induced pressure systems and thus do not display such a west-east pattern (Gacic *et al.*, 1992).

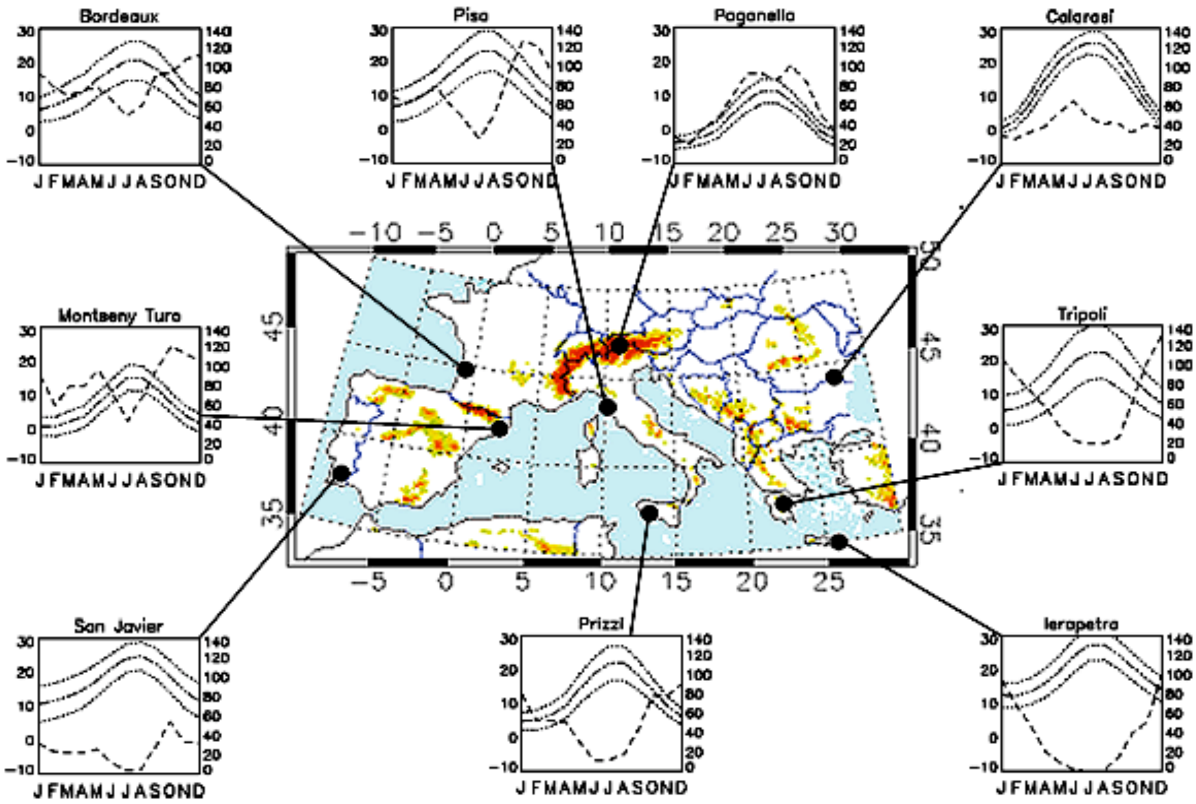


**Figure 2.2: Seasonal profiles (monthly averages) of temperature and precipitation across the Mediterranean. Dashed lines and right axes show total rainfall (mm), dotted lines and left axes show maximum, mean, and minimum temperature (˚C). See Section 3.2.1 for data sources.**

Zonal and meridional patterns are strongest and most polarised in summer and winter, respectively. Although very different from each other the transition seasons are both typically transitive. The autumn season is relatively short with a rapid shift from summer to winter. By contrast, the spring season can be described as 'indecisive' (Barry and Chorley, 1998), often drawn out and possessing indefinite beginning or ending periods. Although the Mediterranean summer has occasionally been defined in an extended fashion (e.g. May-September) (Delitala *et al.*, 2000; Sumner *et al.*, 2001), rainfall re-establishes itself and temperatures start to fall (although slowly) in early September (see Fig 2.2) as incursive circulation strengthens into October (Wallen, 1970). Many Mediterranean countries, particularly to the west (i.e., southern France, Iberia), experience a sudden peak in precipitation in early October (Barry and Chorley, 1998) that signals the collapse of stable summer conditions (see Section 2.2.2). Temperatures drop through October (November in the Aegean), with winter conditions propagating across the majority of the basin (Fig. 2.2), and usually reach a minimum in January, but occasionally in December or February (Lines Escardo, 1977; Furlan, 1977). To fully capture the transition from summer to winter (and vice versa) across the entirety of the basin, the more traditional seasonal definitions of summer as June, July, and August (JJA), autumn as September, October, and November (SON), winter as December, January, and February, and spring as March, April, and May are generally considered appropriate (used by Wallen, 1970; Rodo *et al.*, 1997; Rodriguez-Puebla *et al.*, 1998; Romero *et al.*, 1998a; Agnew and Palutikof, 2000; Gonzalez-Hidalgo *et al.*, 2001, for example) and will be used in this study.

The majority of the northern basin experiences a secondary rainfall peak in March/April (Trewartha, 1961; Barry and Chorley, 1998). In the north and north-east this occasionally exceeds the autumn peak in terms of magnitude (see Fig 2.2). The onset of a definable spring period, this peak is associated with frequent thunderstorms in the west (i.e., Iberia and southern France) and outbreaks of air from beyond the Alps and the Balkan mountain ranges in the north and northeast (Barry and Chorley, 1998). Occasionally also occurring in the south (e.g. SE Spain) the secondary peak is often associated (like the autumn peak) with a southerly air front passing across the region, although it does so in the opposite direction to that in autumn (i.e. northwards) (Goodess, 2000). Discontinuities in topography, the orientation of mountains and the passage of airflow may create an additional one or even two annual peaks in the north-east of the basin (Maheras, 1985), but these are far more localised in nature. Intense outbreaks of rain are one of the typifying characteristics of the Mediterranean region (Wallen, 1970). In the western basin wet, windy, and mild conditions persist through April and May (Wallen, 1970). Rainfall then tails off throughout June, a month that although displaying some spring-like characteristics in the west (Trewartha, 1961), marks the beginning of a warm season that continues until September (Colacino and Conte, 1995). In the east, April is often a dry month, and summer is a more extended period (Barry and Chorely, 1998). In both regions, August has many fewer rainy days than June, temperatures generally peak towards the end of July / beginning of August (Trewartha, 1961). As given in the Köppen classification system (Table 2.1), most of the region experiences less than 30 mm of rainfall in the driest month, less than a third of that in the wettest month. There is, however, more of a discontinuity between the

Mediterranean summer (JJA) and winter (DJF) in terms of both precipitation and temperature in the eastern basin than in the west. The Atlantic provides for more rainfall in summer and more mild conditions in winter, whereas this influence is absent in the east, and outbreaks of northerly air from Europe and Russia can lower winter temperatures considerably (Trewartha, 1961). Mean monthly temperatures as low as –5.7˚C (in Paganella over the period 1958-2000) have been recorded in the north of the basin (Fig. 2.2) in January. Across the Balkan peninsula, the mean values for summer and winter months display remarkable stability: January temperatures are often very similar to February temperatures, and the same is true for July and August (Furlan, 1977).

Having described mean spatial and seasonal patterns across the basin, year-to-year trends in climate are described in the remainder of this section. Compared with the clear global warming trend reported by Working Group 1 (WG1) of the Intergovernmental Panel on Climate Change's (IPCC) Third Assessment Report (TAR), in 2001 (0.6˚C ± 0.2˚C since the end of the 19$^{th}$ century) (see also Mann *et al.*, 1999; Jones *et al.*, 1999a; Mann and Jones, 2003; Jones and Moberg 2003), twentieth century Mediterranean rainfall and temperature trends are less obvious (IPCC, 2001). The TAR states that the most intensive temperature increases have occurred in mid and high latitude Northern Hemisphere continents, and upward temperature trends have been occurring across Europe (Klein-Tank *et al.*, 2002) and the Mediterranean over the last century (Metaxas *et al.*, 1991; Piervitalli *et al.*, 1997; Schönwiese and Rapp, 1997; Piñol *et al.*, 1998; Maheras *et al.*, 1999a; Brunetti *et al.*, 2000a; Quereda Sala *et al.*, 2000; Palmieri *et al.*, 2001; Serra *et al.*, 2001; Baldi *et al.*, 2003). However, temperature changes across the region are far from uniform. Plots presented in the second IPCC assessment report show a general cooling of the Eastern Mediterranean (IPCC, 1995). The SAR plots are supported by both the TAR, and the more recent work of Xoplaki (2002), who shows a contrast between warming in the western and central basin and cooling in the interior of eastern landmasses.

Mediterranean rainfall has been shown to be largely decreasing over the last half century (Folland *et al.*, 1990; Tsonis, 1996; Hurrell and Van Loon, 1997; Schönweise and Rapp, 1997; Zhang *et al.*, 1997; Corte Real *et al.*, 1998a; Piervitalli *et al.*, 1998;

Romero *et al.*, 1998a; Buffoni *et al.*, 1999; Quadrelli *et al.*, 2001; Palutikof, 2003; Xoplaki *et al.*, 2004;). In Italy and Spain, rainfall has reduced by 10-20% over the 1951-1995 period (Piervitalli *et al.*, 1998; Romero *et al.*, 1998a). Over the same period, a decrease of 26% has been found in parts of the southern Mediterranean including southern Italy, southern Spain and Tunisia (Alpert *et al.*, 2002). A comparatively large (winter) downward rainfall trend (up to -2.5mm/decade) has also been found in the southern Adriatic, north-eastern Italy, Portugal, and parts of the eastern Mediterranean (Quadrelli *et al.*, 2001). However, as with temperature, the opposite ends of the basin do not display entirely consistent behaviour (Conte *et al.*, 1989). Some parts of the eastern Mediterranean (particularly southern and central Israel) have been shown to possess an increasing trend in precipitation (Alpert *et al.*, 2002; Palutikof, 2003). Gonzalez-Hidalgo *et al.* (2001) show that precipitation trends can vary over relatively short distances (i.e across the Valencian region of Spain) due to differing geographically related (near-coast, mountainous etc.) mechanisms (see Section 2.2.2). Romero *et al.* (1999a) highlight this localising topographic effect upon rainfall in the south east of Spain. Gonzalez-Hidalgo *et al.* (2001) demonstrated that as with temperature, precipitation trends may vary considerably inter-seasonally. In south-eastern Spain, the majority of stations that show a rainfall trend in winter display significant precipitation decreases, while in spring the majority show increases (Gonzalez-Hidalgo *et al.*, 2001). In north-eastern Italy, a negative precipitation trend is more evident in spring and autumn than winter and summer (Brunetti *et al.*, 2001a). In Turkey, winter rainfall displays an upward trend, in opposition to the summer trend (Kadioglu *et al.* 1999).

### 2.2.2 The Mediterranean setting

The climate characteristics described in the previous section do not occur in isolation but are a reflection of the wider Mediterranean environment, and as such are influenced by a number of important factors. These are principally topography and oceanography, and the air masses that interact with both.

## Topography

Tectonically, the Mediterranean is an example of an inter-continental inter-plate system, and it is the resulting uplift (or more specifically, the Alpine Orogeny) that has produced large mountain ranges (Runcorn 1967) that modify the basin's climate. In the west these ranges include the Spanish Pyrenees, the Cantabrian and the Betic mountains, the Alps, the Italian Appenines, and the Atlas mountains. In the east the Balkan mountain ranges account for roughly 80% of the Balkan Peninsula (Furlan, 1977) and include the Dinaric Alps, the Pindus, the Voras, and the Rhodopes.. As well as the negative (positive) relationship temperature (precipitation) displays with altitude (Agnew and Palutikof, 2000), more dynamical effects exist. Topographic formations act to funnel and intensify, or to shield against, air flow from the north, to obstruct winds from the south or the Atlantic, and to create rain-shadows on their eastern facing flanks (Barry and Chorley, 1998). These effects are partly responsible for the existence of a Mediterranean climate type (Barry and Chorley, 1998), without them the surrounding climatic influences (European, African, Atlantic, Asian) would probably be far more locally prevalent. Spatial variations in topography may create storm conditions or intensify pre-existing systems (Lagouvardos *et al*., 1996). Further, they are conducive to the formation of depressions via the lee-effect that go on to transport water across large stretches of coast line (north-south), or along the length of the basin (west-east) (Barry and Chorley, 1998), as with Genoan cyclogenesis (Bolle, 2003). Cyclogenetic effects are discussed in greater detail in Section 2.3.2. It should be noted that the effects of topography are not limited to large-scale climatology, but are also responsible for microclimatic effects (Milliman *et al*., 1992) that may contribute to the spatially discontinuous trends in temperature and precipitation discussed at the end of Section 2.2.1.

## Oceanography

Two very large bodies of water influence the northern Mediterranean basin: The Atlantic to the west, and the almost entirely enclosed Mediterranean Sea. The Atlantic has two effects upon the region, firstly, it is the source of airflow that carries evaporated water (and relative warmth, during winter months) into the basin (Barry and Chorely, 1998), and secondly it is a major source of inflow for the Mediterranean Sea. Without this inflow the basin would rapidly dry out due to the disproportionate ratio between evaporation and inflow of water due to runoff or direct rainfall (Bigg, 2003; Lolis *et al.*, 2004). Roughly 1600km$^3$/year of water evaporates from the basin, and only 160km$^3$ of runoff replaces it (Groves and Hunt, 1980). This is largely due to the mismatch between incoming and outgoing solar radiation that results in the Mediterranean Sea acting as a heat sink for the majority of the year (Bolle, 2003). The Mediterranean thus provides ample energy and moisture for rainfall systems (Trigo, 2000a) and warms coastal regions during the winter months (Wallen, 1970). Winter low sea surface temperatures (SSTs), around 12ºC, exist in the north east of the basin and the northern Adriatic. The highest summer sea surface temperatures (29ºC) can be found in the north east of the Levantine Sea (Lolis *et al.*, 2004). Much like the orographic uplift effect of topography, oceanically induced conditional instability is a major factor in the creation of a distinct Mediterranean climate (Trewartha, 1961; Lolis *et al.*, 2004). This effect, caused by sensible or latent heat flux between the sea and near surface air, creates buoyancy discontinuity. The heated air rises (i.e., is unstable), and by so doing draws in more air at the surface, creating convergence. This convergence is accompanied by divergence aloft, and by this process the Mediterranean Sea warms and increases the moisture content of incoming air.

Over a period from 1960-1990, Mediterranean Sea level has been decreasing by 0.05mm/year (Schönweise *et al.*,1994; Schönweise and Rapp, 1997). This decrease runs counter to global trends (IPCC, 2001) but has also been detected by Tsimplis and Baker (2000). The change in sea level has been attributed to an increase in Mediterranean Deep Water density (Tsimplis and Baker, 2000), and Mediterranean surface pressure changes (Schönweise *et al.,*1994; Schönweise and Rapp, 1997) driven by the North Atlantic circultion (Tsimplis and Baker, 2000), and an imbalance between evaporation and precipitation rates (Tsimplis *et al.*, 2005). Prior to 1976 Mediterranean Sea Surface temperatures were also decreasing (Reddaway and Bigg, 1996; Cacho *et al.*, 2001). However, during the 1980s this trend displayed a partial reversal west of Greece (Metaxas *et al.*, 1991).

## Air masses

Interacting with both topography and the Mediterranean Sea are discrete air masses, characterised by gradients in pressure driven by local thermal or planetary-scale effects (Ahrens, 1998). Positioned between the European, Asiatic, and African land masses, and the Atlantic Ocean, a number of pressure centres force air over the

Mediterranean mountains and sea. These are subsequently conditioned such that the Mediterranean climate regime described at the beginning of this chapter is created.

To the west, the two centres of action that affect the region more than any other are the coupled pressure systems that occur over Iceland and the Azores (Barry and Chorley, 1998). The Azores high is a 'very large, persistent, zonally elongated area of high sea level pressure' (McIlveen, 1992), a subtropical anticyclone that exists as part of the cellular circulation of the Earth's atmosphere. Conversely, the Icelandic low is a response to convergence at the polar front (Ahrens, 1998), and thus the two systems are linked across the Atlantic Ocean by a rising air mass at ~40-65°N, that sinks between 25 and 35°N. Of interest to this study is the seasonal effect that the Azores high has upon the Mediterranean. During the summer months, it extends eastwards and helps to stabilise the western basin (Trigo and DaCamara, 2000). Later in the year the October peak in precipitation experienced in some regions, southern Spain for example, is a direct response to the withdrawal of the high to the west (Barry and Chorley, 1998), re-introducing Atlantic westerly winds to the region (Bolle, 2003). More detail on this phenomena is given in Section 2.2.3.

To the east, the dominant pressure system is the Asiatic thermal pressure centre (high or low dependent on season). This has a very large spatial extent (Fig 2.3), but is very shallow, and is created by intense heating (during summer) or cooling (winter) of the Asiatic land mass (Ahrens, 1998). This pressure system is characterised by its size and seasonal movement. During the winter season, the Asiatic land mass is at its coldest in Siberia, and this is where the pressure system is centred (as the Siberian thermal high). In summer it becomes a thermal low and moves south and west to occupy the Middle East, the north of India (creating, in part, the 'Asiatic Monsoon'; McIlveen, 1992), north east Africa, and the eastern borders of the Mediterranean basin (Barry and Chorley, 1998; Ahrens, 1998) (see Fig 2.3). The Asiatic summer thermal low is part of the driving mechanism behind the Indian monsoon (Barry and Chorley, 1998) creating landward southerly flow, and produces north-easterly winds across Turkey, Italy and Greece (Ahrens, 1998). A smaller thermal low occurs over Iberia during the summer months, and is described in greater detail in Section 2.3.2.

Shifted northward by the Azores High in summer, the Polar-front upper troposphere jet-stream is the remaining major Mediterranean influence in terms of hemispheric scale air-flow. Part of its eastward flow aloft settles over the Mediterranean in winter, due to a blocking high over Europe at about 20˚W (Barry and Chorley, 1998), and thus exposes the basin to Atlantic storm-tracks (Wallen, 1970). The transport of unstable polar maritime air into the basin by the Polar jet-stream, its lifting (by changes in relief), and subsequent warming (by the Mediterranean Sea) is a major underlying mechanism of both Mediterranean cyclogenesis (see Section 2.3.2, and Trigo *et al.*, 2002) and the creation of the 'Mediterranean air mass' (Barry and Chorley, 1998). Its influence is particularly strong on winter, spring and autumn rainfall (Serrano *et al.*, 1999) and the movement of the Polar jet-stream (coherent aloft, but less so at the surface) helps to characterise the seasonal changeover from meridional to zonal sea level pressure gradients (see Fig 2.3).

**Figure 2.3: Seasonally-averaged (1958-2000) sea level pressure (hPa) derived from NCEP/NCAR reanalysis (Section 3.2.2).**

To the south of Europe, the boundary between the Mediterranean air mass and continental African air creates the 'Mediterranean Front', across which temperatures can vary by 12-16ºC during winter (Barry and Chorley, 1998). Occasionally, when parts of the Icelandic low reach as far south as the north coast of Africa, the Mediterranean Front is altered and warm African air can be drawn from the Sahara into south-eastern Europe (Alpert *et al*., 2005). The resulting hot winds (detailed in Section 2.3.2) may become humid as they cross the Mediterranean Sea (Barry and Chorley, 1998). More generally, infrequent outbreaks of African air cause unseasonal rainfall or exceptional dryness (depending on their path and interaction with the Mediterranean Sea), and high temperatures across the southern parts of Italy, Greece and Turkey (Barry and Chorley, 1998). In addition, a persistent (thermal) low pressure zone over northern Africa may also extend northward and in such circumstances high temperatures may occur across the entire southern boundary of Europe (Giles and Balafoutis, 1990).

### 2.2.3 Oscillations and indices

The air masses described in Section 2.2.2 have wide-ranging effects upon the regions that border them (Hurrell and Van Loon, 1997), and often further afield (Enfield and Meyer, 1997; Rodo *et al.*, 1997; Klein-Tank *et al.*, 1999). To study these effects, and the often oscillatory behaviour of linked air masses (such as the Azores High and the Iceland Low) a number of different indices have gained common usage within climatology, or where no such established indices have previously existed, have recently been developed.

#### The North Atlantic Oscillation

The North Atlantic Oscillation (NAO) is a 'meridional oscillation in atmospheric mass' (Hurrell and Van Loon, 1997) between the two Atlantic pressure systems described in Section 2.2.2, the Azores high and the Iceland low (apparent in Fig. 2.3). In addition to oscillatory behaviour, these pressure centres also exhibit both high and low phases. In a high (low) phase season, differences between the pressure values of the two systems are larger (smaller) than in the opposing phase. Phases are associated with strong zonal air flow (high phase) and weak meridional air flow (low phase), respectively (James, 1995). Due to a pressure contrast across the Atlantic at times exceeding 15mb during high phases, differences in westerly winter windspeeds towards Europe of greater than $8ms^{-1}$ have been recorded between high and low phase seasons (Hurrell, 1995).

The NAO accounts for much of European wintertime temperature and precipitation variability (Hurrell, 1995) and was first measured by Walker (1923) using the pressure difference between the Azores high and the Iceland Low (Van Loon and Rogers, 1978). Defined by the difference between normalised pressures at the southern stations of either Ponta Delgada or Gibraltar, and Iceland, by Jones *et al.* (1997) the North Atlantic Oscillation Index can be calculated. This index exists as a measure of the behaviour of the North Atlantic Oscillation, and is generally also referred to as the NAO. Varying over the last century, from the 1900s to the 1930s the NAO remained high, began to decrease from the 1940s through to the mid 1970s, and then sharply increased into a strong positive phase from 1980 to the mid 1990s (Marshall *et al.*, 2001). More recently the NAO has reduced in strength again (Karcher *et al.*, 2005).

The North Atlantic Oscillation has a positive correlation with both temperature and precipitation across western Europe and the Middle East (Wallace and Gutzler, 1981; Hurrell, 1995; Hurrell and Van Loon 1997; Marshall *et al.*, 2001) and a negative correlation with rainfall across the western Mediterranean and Turkey, strongest during the winter months (Rodo *et al.*, 1997; Esteban Parra *et al.,* 1998; Delitala *et al.*, 2000; Goodess and Jones, 2002; Türkes and Erlat, 2003). The Middle Eastern relationship is mentioned by Hurrell (1995), who describes a decline in moisture transport across the Mediterranean during high phase NAO winters due to the associated north-easterly shifting of the Azores airmass. The Iberian relationship has been found to be purely negative along the Atlantic coast, but positive across some regions of the Mediterranean facing coast (Rodo *et al.*, 1997; Esteban-Parra *et al.*, 1998; Martin *et al.*, 2004). Correlations between southern Mediterranean temperatures or precipitation and the NAO have been found to be largely statistically insignificant (Pozo-Vasquez *et al.*, 2001; Hurrell and Van Loon, 1997) in contrast to the rest of Europe, which demonstrates stronger linkages (Hurrell, 1995; Hurrell and Van Loon, 1997; Rodo *et al.*, 1997; Esteban Para *et al.*, 1998; Delitala *et al.*, 2000; Marshall *et al.*, 2001; Goodess and Jones, 2002). Rather than giving the impression of a lack of influence, however, it has been suggested that the influence of the NAO in this regard is more subtle than elsewhere and mediated by other factors, such as the Mediterranean Oscillation (Maheras *et al.*, 1999a), a phenomenon that is described in Section 2.3.2.

## The All-India Rainfall Index and the Siberian High Index

Significant relationships have also been found between the circulation created by the summer Asian pressure system (see Section 2.2.2 and Fig. 2.3), or rather, the Indian precipitation regime which it controls, and Mediterranean atmospheric and sea level pressure (Baldi *et al.*, 2003; Raichich *et al.*, 2003). Indian precipitation is typified by a weighted index given by the average of rainfall over the entirety of the Indian region during each rainy season (July through September), i.e., the All Indian Rainfall (AIR) index (Robock *et al.*, 2002). Relationships between this and Mediterranean pressure values are shown to be negative across the eastern basin during the Mediterranean July-September dry season, and positive over the western end of the basin during the local September-November precipitation peak (Raichich *et al.*, 2003). Teleconnections have been found between the AIR index and north-west / south-east oscillating Mediterranean Sea level pressure (see Section 2.3.2) (Raichich *et al.*, 2003), the NAO (Robock *et al.* 2003), and pressure variations in the Pacific. Robock *et al.* (2003) found that the AIR index correlated significantly with the winter (DJF) NAO from 1960-1990 and through the 1880s, although the relationship is insignificant in recent years and between 1890 and 1960. In considering relationships between the Asian high and Mediterranean conditions, Ziv *et al.* (2003) offer an alternative to the use of indices, and correlate Asian pressures directly with those of the Mediterranean, resulting in good correlation (r=-0.63) between pressures of the Asian Monsoon and subsidence over the Levant during summer. An intensified summer Asian Monsoon system may create enhanced pressure gradients between it and other nearby systems (such as the east African Monsoon), causing both enhanced subsidence and faster winds (including the Etesian detailed in Table 2.3) (Ziv *et al.*, 2003).

Throughout the winter months, the Asian low is replaced by the Siberian High, which occupies a different position (see Section 2.2.2 and Fig. 2.3), and so cannot be typified by the measures detailed above. The Siberian High Index (SHI) is an area-averaged (40-65°N and 80-120°E) pressure index calculated for 1948-2001 (Panagiatopolous *et al.*, 2005). The region utilized is derived from the study of gridded sea level pressure and station data (covering 1871-2001, combined) across the Siberian region. Currently the index only exists for the winter months (DJF), but significant negative correlations have been found between the SHI and Mediterranean (particularly around Libya) sea level pressure (Panagiatopoulous *et al.*, 2005). Furthermore, positive anomalies in the SHI have been linked to the behaviour of Mediterranean cyclogenesis, and intensification of the Siberian High may be coupled to a strengthening of the jet stream over the eastern Mediterranean Sea (Panagiatopolous *et al.*, 2005). Although no significant correlations have been found with Mediterranean temperature, rainfall in south-eastern Europe is enhanced during periods when the Siberian High is strong (Panagiatopolous *et al.*, 2005).

## North Sea Caspian Pattern

While exploring links between Mediterranean pressure distributions and surface variables, Kutiel *et al.* (2002) found a teleconnection between pressure centers over the North Sea and the Caspian Sea at the 500hPa geopotential height level. These pressure centres have previously been identified as part of the East Atlantic/Western Russia (EAWR) pattern (Horel, 1981; Panagiatopoulous *et al.*, 2002), or the Eurasian (type 2) pattern (Barnston and Livezey, 1987). In the EAWR pattern's negative (more common) phase southerly flow over western Europe is enhanced due to greater than normal geopotential heights over the Caspian Sea, and anomalously low geopotential height values over western Europe and eastern Siberia. Kutiel *et al.* (2002) found that by using a simple spatial differential index similar to the NAO, the North Sea Caspian Pattern Index (NCPI), similar effects could be seen. In its positive (negative) phase the NCPI brings an 'increased (decreased) southwesterly anomaly circulation towards the Balkans, western Turkey, and the Middle East' and increased (decreased) temperatures combined with lowered (increased) precipitation throughout these regions (Kutiel *et al.*, 2002). Thus the NCPI, demonstrating behaviour consistent (though inverse) with the EAWR pattern, can be used as a representation of the interaction of eastern (Eurasian) pressure variation with the eastern Mediterranean basin. Further, Kutiel and Benaroch (2002) have found that the NCPI out-performs both the NAO and indices based upon the Southern Oscillation (detailed below) in terms of differentiating anomalous temperatures (i.e. occurrences of high or low temperature consistently falling during a positive or negative phase) across the Balkans, the Anatolian Peninsula, and the Middle East. The NCPI has no correlation with the NAO but in a similar fashion is generally stronger during winter (Kutiel *et al.*, 2002).

## El Niño Southern Oscillation

In addition to the air masses that border the Mediterranean, previous studies (Rodo *et al.*, 1997; Price *et al.*, 1998; Van Oldenborgh *et al.*, 2000; Lau and Nath, 2001) have shown that surface variables in the Northern Hemisphere respond to the circulation in the South Pacific (Lau and Nath, 2001). The El Niño Southern Oscillation (ENSO) may influence conditions in the North Atlantic and beyond. Price *et al.* (1998) have shown that ENSO may retain its influence as far afield as Israel, enhancing both winter rain and snowfall. This Middle Eastern relationship is only significant over the last 25 years, a period over which the El Niño cycle has significantly increased in intensity and frequency (Trenberth and Hoar, 1997). However, over the period 1931-1990 Turkish rainfall variability and declining December rainfall in the south-east of the country have also both been significantly (at the 0.05 level) related to ENSO (Kadioglu *et al.*, 1999). Karabörk and Kahya (2003) also support a link between a decline in Turkish precipitation and ENSO, particularly with respect to the extreme phases of ENSO.

Further west, Rodo *et al.* (1997) show that while the NAO's influence is strongest in the western part of Iberia, ENSO's power is most evident in the east. Significant positive correlations exist in the east-southeast region of the Iberian peninsula, but not in the west, or the north (Rodo *et al.*, 1997). Rodo and Comin (2000) show that ENSO may have been a powerful forcing agent for south east Iberian rainfall for the last 30 years, and earlier work shows the same kind of effect upon the water levels of 'sensor lakes' since 1910 (Rodo *et al.*, 1997). Piervitali and Colacino (2001), also link ENSO to low levels of precipitation in Sicily.

In Europe, the strongest ENSO effects are subject to a certain amount of lag. Van Oldenborgh *et al.* (2000) found this to be in the range of 3-6 months and almost always strongest in the spring, and then autumn (i.e. for boreal winter SOI values), 'coinciding with the periods of rainfall maxima in temperate areas' of the Iberian peninsula (Rodo *et al*., 1997). Knippertz *et al*. (2003) support this, stating that the highest correlations can be found when considering winter (DJF) ENSO behaviour and following spring (MAM) rainfall. Precipitation is shown to decline across Iberia and northwestern Africa the season after a 'warm' ENSO event. Rodo (2001) reveals a positive correlation between ENSO and cross-Mediterranean cloud cover (+0.4 to +0.6, significant at the 0.01 level) and upper tropospheric relative humidity (up to +0.6, also significant at the 0.01 level), consistent with the relationship found across India and Indonesia, both at a 1-6 month lag. This suggests that Mediterranean climate is linked to ENSO (via the Hadley circulation), and that it may also share connections with the Indian Monsoon, as suggested in the previous section. Price *et al*. (1998) suggest that ENSO may be linked to both conditions in the Atlantic and the Indian Monsoon via shifts in the subtropical jet-stream.

The Southern Oscillation Index (SOI), defined as the normalised pressure difference between Darwin and Tahiti (Ropelewski and Halpert, 1987) is the most commonly used ENSO index, as a direct and comparable counterpart to the NAO index. Many ENSO indices exist, including the Niño3.4 index (an SST anomaly index spanning 5N to 5S and 170-120W) and the MEI (a Multivariate ENSO Index) and each have their uses, but the majority of previous studies concerning southern Europe (including Rodo *et al.*, 1997; Price *et al.*, 1998; Rodo and Comin 2000; Van Oldenborgh *et al.*, 2000) have used the SOI, often in conjunction with the NAO index. The SOI is chosen for its simplicity, availability, and the success with which it represents Pacific conditions (Deser and Wallace, 1987).

## *2.3 Localised dynamics*

### 2.3.1 Localised climatology

Having discussed the effects of atmospheric teleconnection in the previous Section, this Section (2.3) deals with the effects of local atmospheric (spatial) variation upon the Mediterranean climate. When attempting to characterise the behaviour of the atmosphere a number of different variables can be considered. One of the most commonly used factors is sea level pressure (see Section 2.2.3 and below). Others include:

- air and sea surface temperature (Repapis and Philandras, 1988; Hurrel and Van Loon, 1997; Poulos *et al.*, 1997; Cacho *et al.*, 2001; Xoplaki *et al.*, 2003; Xoplaki *et al.*, 2004),
- cloud cover (Piervitalli *et al.*, 1997; Castellari *et al.*, 1998; Trigo *et al.*, 2002b),
- geopotential heights (Kutiel and Kay, 1992; Piervitalli *et al.*, 1997**;** Cavazos, 2000; Karaca *et al.*, 2000; Krichak *et al.*, 2000; Xoplaki *et al.*, 2000; Quadrelli *et al.*, 2001; Türkes *et al.*, 2002; Dunkeloh and Jacobeit, 2003; Xoplaki *et al.*, 2003; Xoplaki *et al.*, 2004),
- specific and relative humidity (Alpert and Neeman, 1992; Tuduri and Ramis, 1997; Eshel and Farrell, 2000; Saaroni and Ziv, 2000),
- and zonal and meridional components of wind velocity (Piervitalli *et al.*, 1997; Poulos *et al.*, 1997; Tuduri and Ramis, 1997**)**.

Each of the above continuous variables may help to represent local horizontal and vertical energy and moisture transport (see below), or discrete phenomena (Section 2.3.2). In this study three of the above factors (important to mesoscale circulation) are considered in greater detail (Chapter 3), and are discussed further below. Average fields for these factors (over a period from 1958-2000) are illustrated in Figures 2.3 (Section 2.2.2), 2.4, and 2.5.

## Sea level pressure (SLP)

Changes in sea level pressure reflect the transport of energy (i.e. heat) from one region to another, and can be used to describe global atmospheric circulation in terms of high and low pressure centres (e.g. the Azores High and the Iceland Low, or the Asiatic thermal system). In regional weather forecasting, SLP readings are taken to map surface level features (Fig. 2.3). This process facilitates the prediction of both localised wind flow (Section 2.3.2) between areas of high and low surface pressure (Pirazzoli and Tomasin, 2003; Pirazzoli, 2005) and other frontal, cyclonic, or anticyclonic weather (Ahrens, 1998). The majority of air masses and oscillation indices detailed in Sections

2.2.2 and 2.2.3 are described using SLP values and in climatology specific pressure patterns have been linked to dry or wet conditions across the Mediterranean (Kutiel and Paz, 1998; Chessa *et al.*, 1997; Krichak *et al.*, 2000). Numerous studies have utilised the association between SLP and surface weather to categorise different Mediterranean circulation 'types' (i.e. distributions of high and low pressure) that affect temperature, rainfall, or both (Section 2.3.3).

## Geo-potential height (Z)

The geopotential height of a fixed pressure level is a measure of the volume of air beneath that level in terms of altitude, where height is redefined as an expression of work against gravity (Ahrens, 1998; McIlveen, 1992). Given a fixed pressure level, this quantity varies with both the mass of air beneath the relevant altitude and the temperature of the given air column. For this reason, the geopotential height of a fixed pressure level generally decreases with increasing latitude (Fig. 2.4).

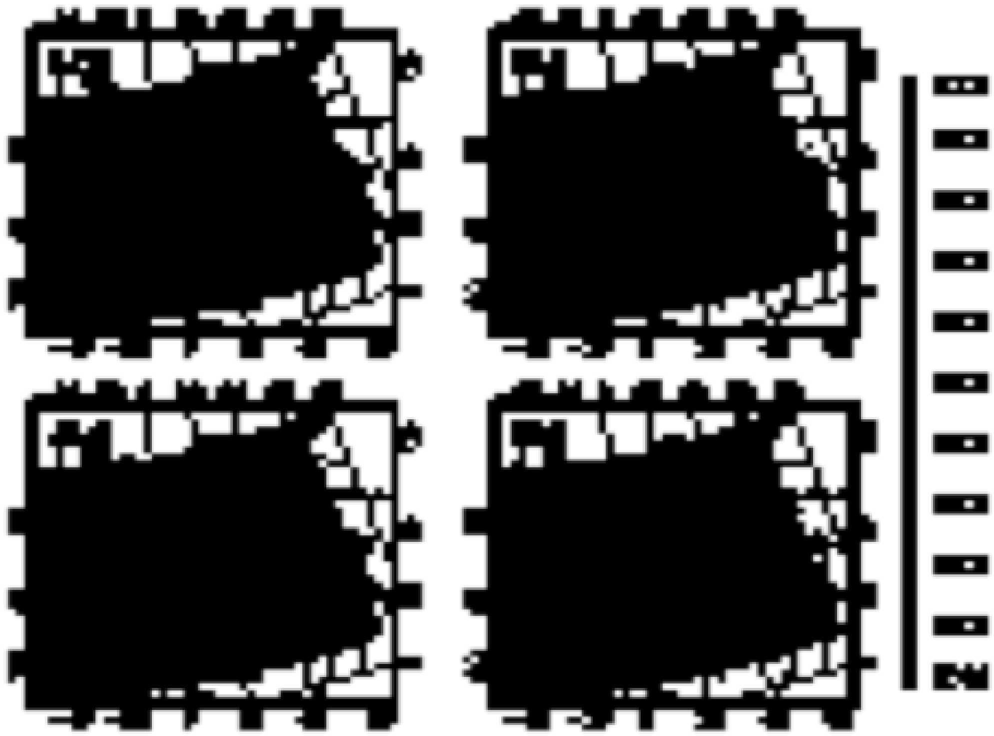

**Figure 2.4: Seasonally averaged (1958-2000) geopotential heights at the 500hPa level (in metres) derived from NCEP/NCAR reanalysis.**

Due to a temperature differential between land and sea, winter contours over land are shifted southward, and are wider than over the Atlantic, creating divergence and weakening air speed aloft to give more intense surface high pressure zones over water than on land (Ahrens, 1998). During spring the average field is a weakened form of the winter field, reflecting smaller temperature contrasts and weakened flow. Values shift northward due to warmer conditions, particularly in the east of the region. Average summer geopotential heights show a reversed contrast between land and sea temperatures from winter (surface high pressure systems more intense over land than sea), and a very large high over west Saharan Africa. Autumn contours are very similar to those for spring, although Z500 values are generally higher for a given region. This is particularly the case over northern Europe (Fig. 2.4).

Map values of geopotential height over a given area, which show departures from the simple latitudinal relationship due to variations in land/sea mass and temperature as given above, can be used in addition to SLP maps to represent cyclones and anticyclones, to calculate geostrophic wind, and to otherwise give an estimate of localised airflow (Ahrens, 1998). Geopotental height patterns have also been linked to temperature variations (Xoplaki *et al.*, 2003) and wet or dry conditions in the Mediterranean (Cavazos, 2000; Krichak *et al.*, 2000; Xoplaki *et al.*, 2000; Turkes *et al.*, 2002; Dunkeloh and Jacobeit, 2003; Xoplaki *et al.*, 2004). In addition to SLP data, circulation schemes (Section 2.3.3) have been constructed from geopotential height data for the region and linked to both temperature and precipitation.

Different geopotential height levels have different applications within climatology. For this study the 500 hPa (mid-troposphere) level is under consideration, as the movement and speed of large scale surface weather conditions are linked to the direction and speed of 500 hPa wind flow (Ahrens, 1998). In weather forecasting, 500 hPa charts are used to gain a picture of whether surface conditions (including cyclones) are strengthening or weakening due to divergence (difluence) or convergence (confluence) at height. Extra-tropical storms of the kind active in both the North Atlantic and the Mediterranean basin may be characterised with the use of 500 hPa features (Beersma *et al.*, 1997; Carnell and Senior, 1998; Karaca *et al.*, 2000).

## Specific humidity (SHM)

Specific humidity is the ratio of water vapour mass to the total mass of air in a moist air parcel. More than 90% of atmospheric moisture lies below the 500hPa level, 50% below 850hPa (Barry and Chorley, 1998), due to this distribution both cloud and precipitation are largely confined to the troposphere. Sources of moisture for the Mediterranean include the Mediterranean Sea (Bigg, 2003), the Atlantic, and the Tropics. Tropical plumes can transport moisture from highly humid regions (monsoonal Asia and equatorial Africa) into the Mediterranean around the east of the Sahara (Fig. 2.5), a process important for the generation of rainfall in arid regions (Rubin *et al.*, 2005).

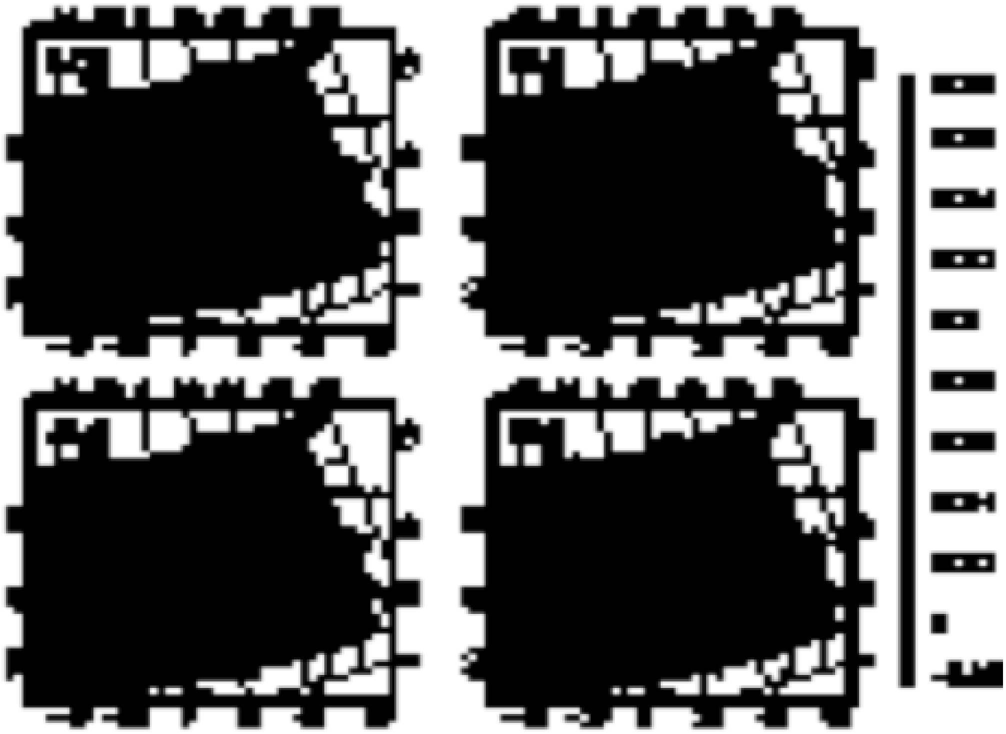

**Figure 2.5: Seasonally averaged (1958-2000) specific humidity at the 850mb level (g/kg) derived from NCEP/NCAR reanalysis.**

The principal factors acting on the atmospheric moisture balance (in the Eastern Mediterranean) are rainfall and the vertical transport (below 850mb) of water vapour (Eshel and Farrell, 2000). The majority of precipitation forms at the 850mb level, and a strong relationship (independent of topography) exists between 850mb specific humidity and the majority of precipitable water (Ross and Elliott, 2000). Due to this relationship maps of 850mb specific humidity can be utilised to study the transportation of water vapour utilised in rain and storm generation. The strength of upward moisture advection (i.e. the amount of water vapour at height) has a direct influence upon the intensity of both rainfall and convection, and can be used to explain heavy episodes of precipitation (Tuduri and Ramis, 1997; Eshel and Farrell, 2000). High levels of humidity throughout the troposphere may help to trigger particularly heavy episodes of convective rainfall (Tuduri and Ramis, 1997).

Of the factors listed at the beginning of this section, wind components have been neglected in this study due to their strong dependence on pressure gradient force (Ahrens, 1998). As both pressures at the surface level and at height are considered, and the extremes detailed in this study (which do not include wind storms) exist at the mesoscale and above, wind components would represent an intermediary, ancillary level of detail. Although links have been shown between cloud cover, pressure, and rainfall (Piervitalli *et al.*, 1997), and total cloudiness is known to modify daily temperature ranges in association with atmospheric circulation (IPCC, 2001; Trigo *et al.,* 2002a), complicated feedbacks exist between temperature and cloud cover, and different types of clouds may have different effects upon surface climatology (IPCC, 2001). Although cloudiness (like wind speed) may act as an intermediary factor between mesoscale circulation and extremes of temperature and precipitation, to accurately represent the wide range of processes cloud cover may modify is beyond the scope of this thesis. Studies into basin-wide ocean-atmosphere coupling beyond buoyancy transfer or momentum flux (the movement of surface water via the motion of winds) are largely inconclusive to date, although Marrachi *et al.* (2000) have found correlations between September and June SSTs and October and December extreme rainfall. Further afield, Rowell (2003) has shown that Mediterranean SSTs may affect Sahelian (central Africa) precipitation during the African rainy season. However, Xoplaki *et al.* (2004) found no improvement in model skill when using SST as a predictor (in addition to SLP and geopotential heights) for Mediterranean rainfall and little skill when utilised as the only predictor. They conclude that SSTs are only indirectly linked to Mediterranean rainfall, and are forced by factors evident in both sea level pressure and geopotential heights. Lolis *et al.* (2002) show a link between lower troposphere temperatures and SSTs, but it is the former that leads the latter at a one-month lag.

Extreme climate events are often characterised by unusual features in local SLP fields (Lagouvardos *et al.*, 1996; Krichak *et al.*, 2000; Trigo and Davies, 2002; Tuduri *et al.*, 2003; Esteban *et al.,* 2005), Z500 (Llasat and Puigcerver, 1994; Chessa *et al.*, 1997; Sioutas and Flocas, 2003; Anagnostopoulou *et al.*, 2006; Prezerakos *et al.*, 2006) or SHM (Tuduri and Ramis, 1997; Eshel and Farell, 2000; Saaroni and Ziv, 2000). Very high geopotential heights generally describe strong mid-troposphere ridges and particularly calm/warm surface conditions. Very low 500 hPa values may represent an intense mid-troposphere trough or surface-level storm. SHM values may be (when low) associated with the persistence of hot and dry conditions (Saaroni and Ziv, 2000), or (when high) the transport of large volumes of water for intense rainfall (Tuduri and Ramis, 1997**)**.

Links between extreme events and local circulation are discussed further in Section 2.3.3 but coherent features that can be found in Mediterranean SLP, Z500 and SHM fields, and that may have an affect upon extreme Mediterranean climatology are discussed below, including:

- Oscillations caused by atmospheric waves (Z500),
- Thermally induced high or low pressure zones (SLP),
- Travelling regions of intense low pressure and relatively high vorticity (SLP, Z500, SHM),
- Strong (named) winds generated by opposing pressure centres (SLP, SHM),
- And circulation patterns associated with given weather regimes (SLP, Z500, SHM).

### 2.3.2 Dynamic phenomena

## The Mediterranean Oscillation

Although some discussion continues as to whether it is a localised oscillation, or an extension of the NAO (Palutikof *et al.*,1996; Maheras *et al.*, 1999a), a substantial body of work indicates the existence of an oscillating atmospheric pressure differential across the Mediterranean basin (Corte-Real *et al.*, 1995; Corte-Real *et al.*, 1998b; Kutiel and Paz, 1998; Maheras *et al.*, 1999b; Piervitalli *et al.*, 1999; Baldi *et al.*, 2003). Centres of action have been taken at opposing ends of the region, between Algiers and Cairo (as in Conte *et al.*, 1989; Palutikof *et al.*, 1996; Corte-Real *et al.*, 1998b; Piervitalli *et al.*, 1999), or Gibraltar and Israel (North Front, and Lod airport stations respectively) (Palutikof *et al.*, 2003). Throughout this study 'Mediterranean Oscillation' (or MO) is taken to refer to the phenomenon itself, and the two (highly inter-correlated) indices produced from the different centres of action (defined by 500hPa level geopotential heights) are referred to as MOI(AC) and MOI(GI). Differing Z500 values at either end of the basin can be seen for both winter and summer in Fig.2.4.

Geopotential height (500hPa) anomalies at the two ends of the basin trend towards zero in a periodic fashion, such that values at the western (eastern) end increase (decrease) over time with a 22 year cycle in anti-phase with the opposing centre (Palutikof *et al.*, 1996; Corte Real *et al.*, 1998b; Piervitalli *et al.*, 1999). This 'seesaw' motion is the typifying behaviour of the oscillation (Maheras *et al.*, 1999a): upward pressure trends at one end of the basin occur in synchrony with downward trends at the other. Palutikof *et al.* (1996) suggest that this motion is an expression of 'long atmospheric waves', half the wavelength of which (1500km) approximates the length of the basin.

Corte-Real *et al.* (1998b) suggest that the positive western pressure trend is connected to an increased persistence of the Azores High over that part of the basin, and that this anticyclonic system should inhibit precipitation. Palutikof (2003) concludes that the MOI(GI) has a significant negative relationship with rainfall over the central and northern Mediterranean in summer and winter months, over the western Mediterranean during winter and autumn, and over the northern region in all seasons except spring. Through analysing correlations and standardized anomaly indices (SAIs- a method of compositing station data) she finds that during summer and throughout the central and northern Mediterranean the MOI(GI) is observed as an independent entity, while for the rest of the year and region it represents more of an extension of the NAO than an independent oscillation. Piervitalli *et al.* (1999) demonstrate that the eastern Iberian and Italian regions show the highest negative correlations between MO and rainfall, exceeding values for those associated with the NAO throughout the year.

Maheras and Kutiel (1999) describe temperature conditions at either end of the basin (eastern: east of 20E, and western: west of 10E) as consistent with the oscillating Mediterranean pressure regime. Southerly circulation at one end of the basin has been associated with northerly conditions at the other, bringing unusually high or low temperatures respectively. Maheras *et al.* (1999b) provide further detail concerning the relationship. Low eastern pressures result in an invasion of African air over the central basin, and westerly flow across the Balkans. Such a regime produces high temperatures throughout the Mediterranean. In the opposite phase, low pressures exist at the westerly end of the basin and draw colder air down from Europe.

The study of Corte-Real *et al.* (1995) suggests that the Mediterranean Oscillation produces a north-west / south-east pressure gradient, rather than simply east-west, similar to the gradient produced by the North Sea Caspian pressure pattern. Kutiel and Benaroch (2002) intended that the NCPI should replace pre-existing Mediterranean Oscillation indices, and the phases of the MO and NSCP bring similar temperature effects. However, the indices associated with these patterns do not correlate well (Section 3.5.3), which suggests that they may not be directly analogous representations of the same phenomenon.

A further representation of oscillating north-west / south-east Mediterranean (sea level) pressures between Mersa Matruh (about 240km west of Alexandria) and Marseille has been created by Raichich *et al.* (2003). Significantly correlated with both Indian rainfall (–0.68) and African (Sahelian) rainfall (–0.47) this configuration suggests that Mediterranean indices may provide useful linkages with both western and eastern circulation. However, the Raichich *et al.* (2003) study concerns only such distant linkages and not the effects that an east/west oscillation may have on Mediterranean climate.

## The Iberian thermal low

Similar in nature to the Asiatic thermal low, but on a much smaller scale, summer heating of the Iberian peninsula causes a surface pressure low (Alonso *et al.*, 1994), visible in both Fig. 2.3, and as a deformation of contours in Fig. 2.4. This depression then creates a counter-clockwise flow over the peninsula (Lines Escardo, 1977). The Iberian thermal low has been linked to hot and dry winds from Morocco, when the African thermal low joins with that of Iberia, helping to enforce dry and stable conditions across Spain and Portugal (Stanislawski, 1959). Stable conditions can persist for some time (in the order of weeks), as the daily summer warming of the peninsula may not be entirely removed by the usual night-time cooling (Millán *et al.*, 1997), leaving a residual low at the beginning of each day (McIlveen, 1992). Enhanced convergence and ascent in the centre of the peninsula also intensifies winds over the complex terrain of the Iberian interior (Homar *et al.*, 2003; Pérez *et al.*, 2004). If the Iberian thermal low draws air from the Atlantic, rather than Africa, incursive winds can become humid, conditions become unstable, and intense thunderstorms may occur (Lines Escardo, 1977).

## Cyclogenesis

One of the principal forms of control for intense rain generation and moisture transport (Fig. 2.5) in the Mediterranean is the passage of low pressure systems (i.e. depressions, or cyclones) across the region. Cyclogenesis is particularly important for winter rainfall across the basin, from the Iberian peninsula (Trigo *et al.,* 2000a) to Greece. Northern and western winter Greek rainfall is, in fact, mainly controlled by cyclogenesis that occurs in the Gulf of Genoa, many miles away (Bartzokas *et al.*, 2003). Of the total number of cyclones present within the basin only ~40% are persistent (lasting longer than 12 hours, and on average possessing a 28 hour lifespan), and ~35% are large: over 550km in radius (Trigo, 2000a) The remainder consist of a large number of 'background' small, short lived, systems (Barry and Chorley, 1998). The majority of large cyclones move from west to east (Wallen, 1970). They are formed in definite regions within the Mediterranean (such as the Gulf of Genoa), each with their own characteristic qualities, as summarised in Table 2.2. Winter depressions in sea level pressure also form directly over tthe Adriatic, the Ionian Sea and Crete.

In each case, cyclones are formed in association with the effects of complex topography and the Mediterranean Sea. They are often subsequently steered by the former, and reinforced by the latter (Trigo, 2000a). Dynamically unstable flow aloft is common to most forms of Mediterranean cyclogenesis, and an upper-level trough traversing the region can trigger successive depressions, aided by such effects as lee-effect cyclogenesis and the re-intensification of older cyclones (Trigo, 2000a). Successive divergence and convergence of air causes the first of these phenomena, as it is squeezed over a topographic barrier and then released at a lower altitude and latitude. This is due in turn to the vertical incompressibility of insurgent flow, and residual cyclonic relative vorticity from the decrease in planetary vorticity from higher latitudes to lower (McIlveen, 1992). The two predominant Mediterranean centres of lee-effect cyclogenesis (North Africa and the Gulf of Genoa) are thus to the south-east of substantial mountain ranges, i.e. the Atlases (Fig. 2.5) and the Alps (Trigo, 2000a). The most intense Mediterranean cyclones are generally produced within these regions (Trigo *et al.*, 2000a). Re-intensification effects are partially lee-effect (e.g. around Cyprus) and partially thermal in origin, due to convective instability above the Mediterranean Sea (particularly in the eastern basin, south of Greece and Turkey). Requiring localised temperature contrasts, the regenerative thermal process is more evident in winter, when the Mediterranean is not so uniformly heated. This, and the influence of strengthening (eastern) or weakening (western) hemispheric scale circulation (e.g. the North Atlantic Oscillation and the jet-stream) means that cyclogenesis is usually weaker in summer than in winter, and many of the tracks (particularly in the central and eastern Mediterranean) disappear in the warmer months, becoming quasi-stationary (Alpert *et al.*, 1990) and losing their ability to transfer moisture. As roughly 60% of cyclones in the Mediterranean Basin last less than 12 hours the majority do not leave the Mediterranean, but die within the region. Not all the cyclones that influence the Mediterranean basin develop within the region. Roughly 9% of cyclones influencing surface conditions in the Mediterranean are intruding Atlantic cyclones (Barry and Chorley, 1998). Formed outside of the basin, they are a result of the Atlantic front between cold polar air and tropical maritime air (Barry and Chorley, 1998).

Although it has been found that intense cyclones are increasing in frequency in the Northern Hemisphere, for winter the increase is largely for latitudes over 60°N (Lambert, 1995; Serreze *et al.*, 1997; Key and Chan *et al.,* 1999). For the 1958-1997 period and between 30-60°N a significant decrease (-0.50, significant at the 0.10 level) has been detected for the total number of cyclones (Key and Chan *et al.,* 1999; McCabe *et al.*, 2001) and the frequency of intense winter cyclones across the Mediterranean is in decline (Piervitalli *et al.*, 1997; Trigo *et al.*, 2000a). This change in mid-latitude frequency has been linked to increases in winter temperature (Serreze *et al.*, 1997), significantly correlated at –0.58 (McCabe *et al.*, 2001) and, indirectly, changes in North Atlantic climate variability (Trigo *et al.*, 2000a). McCabe *et al.* (2001) found that the trend in mid-latitude cyclones reflects a mid 1970s North Atlantic regime shift (Marshall *et al.,* 2001). However, although the frequency of mid-latitude intense cyclones is in decline, the intensity of cyclones occurring in both high-latitudes and mid-latitudes is increasing significantly (McCabe *et al.* 2001), if in a manner unrelated to temperature (no significant correlations).

Declining rainfall has been attributed to an increase in anticyclonic behaviour in both the western and eastern ends of the Mediterranean basin (Bartzokas *et al.*, 2003). Trigo *et al.*, (2000a) have linked a decline in extended Northern Mediterranean winter rainfall (monthly rainfall values between October and March over 7.5°W-45°E, 37.5-50°N) with a decrease in the frequency of cyclones. Cyclonic activity can explain 72% of the variance of extended winter rainfall, which correlates significantly (at the 5% level) with the frequency of both intense (0.68) and non-intense (0.6) cyclones (Trigo *et al.*, 2000a). Jansa *et al.* (1996; 2001) have found that even small and short-lived cyclones may have an indirect effect upon intense rainfall.

## Mediterranean winds

A number of regional winds are well known across the Mediterranean basin for their distinctive characteristics. These are summarised in Table 2.3 and are largely caused by airflow between the pressure phenomena detailed above and in Section 2.2. Many of the winds bring either unusually hot (e.g. the Scirocco) or cold (e.g. the Bize) conditions to their particular regions, and may bear intense rainfall (e.g. the Gregale), while a few bring stable weather conditions instead (e.g. the Mistral).

Piervitalli *et al.* (1997) have found that strong (over 46 $kmh^{-1}$) wind events for the Mistral (in Capo Frasca, Sardinia), the Bora (in Trieste), and the Scirocco (in Trapani, Sicily) have all been decreasing in frequency (between 1951-1990), with a sharp decline since the 1980s. Pirazzoli and Tomasin (1999) also found a decrease in the frequency and strength of the Bora, among other easterly winds. In contrast with wind activity on the western coast (Pirazzoli, 2005), southerly wind frequencies along the southern coast of France have been (slightly) decreasing since 1951 (Pirazzoli, *et al.,* 2004). The general decline in central Mediterranean wind frequency and velocity has been linked to an increase in pressure (at the surface level and at the 500hPa level) over the west and central Mediterranean basin (Piervitalli *et al.* 1997), consistent with the decrease in cyclonic activity detailed above. However, in another study (Pirazzoli and Tomasin, 2003), a decrease in frequency and velocity for Italian coastal winds was found only for the period 1951 to the mid 1970s. Toward the end of the 20$^{st}$ Century an increase in both wind frequency and velocity for Italian coastal stations (Pirazzoli and Tomasin, 2003) has been linked to temperature (through similar trends).

### 2.3.3 Circulation typing

The above sections show that the short time scale climate (i.e., over days and weeks) of the Mediterranean is highly dependent upon localised circulation. Such circulation can be a smaller scale product of hemispheric control (as with the MO or the NSCP) or the result of 'internal' processes, including cyclogenesis, differential heating and topography. A significant proportion of Mediterranean climate research is given over to classifying past circulatory conditions as an aid to understanding how changing atmospheric circulation produces specific surface weather conditions (Corte-Real *et al.*, 1995; Zhang *et al.*, 1997; Goodess and Palutikof, 1998; Maheras *et al.*, 1999a; Goodess and Jones, 2002). Also of interest to a number of studies (Zhang *et al.*, 1997; Maheras *et al.*, 1999a; Goodess and Jones, 2002) are the trends apparent in the frequencies of different types of circulation, as an aid to understanding the explicit mechanisms that force surface conditions over time (Zhang *et al.*, 1997). Circulation typing is largely pursued through either of two approaches; manual classification, or automatic classification. Manual classification is self-descriptive, and involves the division of surface and upper-air circulation maps over a target region into groupings using expert judgement. Usually these classifications are defined by the degree and position of cyclonicity or anti-cyclonicity, or upon a dominant direction of flow, and reflect surface conditions in a largely consistent manner. Automatic classification is essentially the same process, but utilises recent improvements in computational processing power to apply more mathematically consistent logic. The latter is dependent upon computational rules, and is thus more rigorous, more sophisticated, and reproducible. The former has a greater volume of supporting literature (Baur, 1947; Del Trono, 1965; Hess and Brezowsky, 1969; Karalis, 1969; Pinna, 1970; Lamb 1972; Hess and Brezowsky, 1977; Maheras, 1979; Lines Escardo, 1981; Comrie, 1992; Sweeney and O'Hare, 1992; Leathers and Ellis, 1996; Yarnal and Frakes, 1997; Zelenka, 1997; Prezerakos, 1998; Weber, 1998; Yarnal *et al.*, 2001), is more flexible, and allows for a greater degree of informed insight, but ultimately is more subjective, and more time consuming (Yarnal *et al.*, 2001; Kostopoulou, 2003).

The most famous of the European manual classification systems are those conceived of by Baur (1947) and Lamb (1972). The European Grosswetterlagen scheme of Baur, Hess, and Brezowsky (Hess and Brezowsky, 1969; Hess and Brezowsky, 1977) is dependent upon persistence (3 days or more) of circulation, as well as form (i.e. zonal, meridional, or mixed). Grosswetterlagen were initially applied to a region encompassing the eastern North Atlantic and Europe to capture atmospheric flow (surface level and at height) at the synoptic scale. In addition to the three general groupings above, a further 10 major, and 29 sub-types (plus one unclassifiable type) have been defined according to position of pressure centres (e.g. north, west, central European, etc.) and position (e.g. British Islands, Fennoscandian, Norwegian…) combined with type (e.g. cyclonic or anticyclonic, ridge or trough). The Lamb system is also two tier, defining flow as meridional, zonal or mixed type, before dividing circulation further into specific cases (e.g. North, Northeast, Cyclonic, Anticyclonic, Hybrid Cyclonic…) (Goodess and Palutikof, 1998). Lamb's classification was formulated for the British Isles, but can be applied to any northern mid-latitude location, and utilised forms of pressure system evolution to define classes of weather, and hybrid types to represent mixed conditions. Like Baur, Hess, and Brezowsky, Lamb recognised that he was unable to classify weak conditions or those that changed quickly, and thus his classifications also include an unclassifiable division. More recent attempts at classification are largely improvements or adaptations of these two approaches.

Recent improvements to the manual method include those by Comrie (1992), and Frakes and Yarnal (1997). The first of these utilised a form of statistical analysis ('declimatizing') and the frequencies of pre-defined weather types to distinguish between synoptic and non-synoptic components of environmental data (e.g. visibility). The second introduced correlation analysis to speed up the, otherwise slow, manual method (Yarnal *et al.*, 2001). The Lamb classification system has largely been replaced by an automated scheme, the Jenkinson and Collison method (Jones *et al.*, 1993), but other manual methods have continued to be used. Yarnal *et al.* (2001) have reviewed manual classification applications for the 1990s, and describe work that has utilised manual typing with a wide range of parameters (rainfall, snowfall, temperature, wind flow, ozone levels, stream flow, and aerosol concentrations) in addition to circulation (sea level pressure and 500 hPa geopotential heights). Mediterranean weather type classification using the manual method remains largely limited to circulation (Del Trono, 1965; Karalis, 1969; Pinna, 1970; Maheras, 1979), temperature (Prezerakos, 1998), and precipitation data (Lines Escardo, 1981; Sioutas and Flocas, 2003).

Automated classification relies upon a variety of statistical techniques, from the relatively simple (correlational analyses), e.g. Lund (1963), to the more complex (neural networks and other forms of artificial intelligence), e.g Trigo and Palutikof (1999), Cavazos (2000), and Trigo (2000b). The majority of classification systems relevant to this work are of the eigenvector type (Lorenz, 1956), specifically utilising a factor analysis technique called 'principal components analysis' (PCA) (Preisendorfer, 1988; Wilks, 1995). Eigenvector based work is both mathematically compact and statistically robust (Kostopoulou, 2003) and among the varying kinds (e.g. cluster analysis, canonical correlation analysis, discriminant analysis) PCA tends to provide a greater level of information than those solely dependent on empirical orthogonal functions (Yarnal, 1993; Yarnal *et al.*, 2001). PCA utilises either covariance or correlation matrices to find components of variance, patterns which may or may not represent underlying driving factors (Wilks, 1995). Further, the components of such an analysis are often rotated to simplify the distribution of loadings and aid in interpretation

(Bloomfield and Davis, 1994, Mestas-Nuñez, 2000). However, rotation of principal components may remove useful properties (successive explanation of variance, orthogonality, and minimal cross correlation) of un-rotated components (Yarnal, 1993). Further detail regarding this form of statistical variance exploration is provided in Chapter 3.3.6. As with manual classification, a wide range of non-circulatory applications exist (classifying rainfall, temperature, air quality, surge events, and severe storms) in addition to those focused upon pressure based circulation typing, examples of which are also given by Yarnal *et al*. (2001). Examples of available Mediterranean automated classification literature, applied to circulation types, are summarised in Table 2.4.

### 2.3.4 Hemispheric / local coupling

The preceding sections on Mediterranean climatological processes build a picture of a highly inter-related system that operates on a number of different spatial scales. Maheras *et al.* (1999b) state that:

'The Mediterranean can be regarded as a transitional zone between the continental influences of Europe and Asia, the desert climate of North Africa and the oceanic effects from the Atlantic. The surface pressure field is influenced by the oceanic Azores High, the Siberian winter anticyclone, the northwestern extension of the South Asian thermal low, as well as more transient anticyclones and travelling depressions'

This quote neatly summarises the previous sections of this chapter. Hemispheric-scale conditions affecting regional surface variables (such as temperature and precipitation) exert their influence on local atmospheric circulation, and the propensity of certain regions to produce depressions. In turn these cyclones bring rain across the region, largely laterally from west to east on a Mediterranean scale, and rotationally across the south-western coasts of countries on a national scale. They also bring with them characteristic winds, through which the climate of one region extends to influence another (as with the Scirocco, rising in Africa and the Near East to influence the central Mediterranean). Both winds and cyclones may be modified during their lifespan, by the relative heating (cooling) influence of the winter (summer) Mediterranean Sea, or the influence of topography, which can either intensify or divert flow.

This then is a coupled system in which each scale of phenomenon interacts with the next. Examples of these couplings, summarising previous sections, are given here:

- Topographic effects that intensify flow are partially responsible for a number of the named winds that characterise and directly modify the surface climate of some parts of the basin (see Table 2.3). Other winds are generally the result of Mediterranean cyclogenesis, a phenomenon also aided by topographic processes, e.g., the lee effect (Barry and Chorley, 1998).
- The work of Trigo (2000a) shows that sensible and latent ocean/atmosphere heat flux is concentrated in regions of cyclogenesis (i.e., the Gulf Du Lyon, the Gulf of Genoa, and the near-Mediterranean Atlantic) during months of intense storminess, and is at a minimum during summer, as with cyclone activity. This oceanic/cyclogenetic link demonstrates the effect that energy transport via the Mediterranean Sea has upon the region.
- Alpert *et al.* (1990) have shown that the positions of centres of cyclogenesis change seasonally due to interactions between topography and oceanography altering the direction of airflow. Month to month variations in the tracks of resultant depressions may be similarly dependent.
- The Azores high can expand as far as the eastern Mediterranean during winter, deflecting the polar jet stream, and thus decreasing the power of local cyclones. This is far from an annual occurrence, but is partially responsible for declining rainfall in some regions (Trigo and Davies, 2000) and has been linked to the behaviour of the MO (Palutikof, 1996).
- The west-east paths of cyclones across the Mediterranean basin (from the Atlantic, the Gulf of Genoa, or Northern Africa) are associated with the jet stream and are stopped or diverted (depending on season) in their easterly extent by the Asiatic pressure system (Alpert *et al.*, 1990; Barry and Chorley, 1998). The Iberian low has been linked with weakened Algerian Mediterranean cyclogenesis during the summer season (Alonso *et al.*, 1994), and therefore helps to modify conditions as far afield as Greece (Egger *et al.*, 1995).

- Ziv *et al.* (2003) have demonstrated linkages (lagged at 2 days) between weakening descent of air in the eastern basin and the Asian pressure regime. The resultant, less stable, weather may then influence both rainfall and temperature in the eastern Mediterranean region.
- Meridional circulation over the Balkans has been related to an increase in precipitation, while the opposite is true for zonal circulation (Maheras and Kolyva-Machera, 1990). Also, Maheras *et al.* (1999a) have linked persistent anticyclonic conditions over Greece to decreasing precipitation. Both findings illustrate a plausible link with Asiatic pressure variations.
- The results of Maheras *et al.*, (1999b) show definite influences upon temperature by the Atlantic in the west Mediterranean and a combination of southern Asian, north African and central European circulation influence in the east.

Thus, linkages between large-scale effects and more localised patterns of Mediterranean surface temperature and precipitation can be demonstrated via the medium of synoptic-scale circulation (Bartzokas and Metaxas, 1991; Corte-Real *et al.*, 1995). Also, the usage of indicators of hemispheric circulation to explore such relationships can be justified (Maheras and Kutiel, 1999). We can therefore validly theorise about relationships between large-scale effects and the extremes of surface variable distributions. An example of such a relationship is the 22 yr period of the Mediterranean Oscillation and the similar periodicity associated with Mediterranean heatwaves (Colacino and Conte, 1995). Extremes of climate and the way they relate to atmospheric circulation are explored in the next section (Section 2.4).

## 2.4 Climatological extremes across the Mediterranean

### 2.4.1 Extremes as a part of Mediterranean climate

In contrast to the mean monthly temperatures shown in Figure 2.2, across the Northern Mediterranean literature from before 1980 (Lines Escardo, 1970) reports maximum temperatures of 50˚C in the west (Alto Douro, Portugal), with minima as low as –30˚C at inhabited altitudes (905m, Calamocha, Spain). In the eastern basin, minimum temperatures were reported in shallow basins within the lower Alps (-32.4˚C), and in the lowlands of Romania and Bulgaria (below –30˚C). The greatest eastern basin daily maximum temperatures were found to approach 45˚C in Larisa and Bucharest. Maximum daily temperatures for the whole of the target region were reported on the southern border, in Algeria and Libya, where temperatures exceeded 50˚C (57˚C in Al Azizia, Libya; Griffiths, 1972) during the height of summer.

More recent records (i.e. daily station data for the target region for the period 1960-1999, see Section 3.2.1) provided by the Fundación para la Investigación del Clima (FIC) show a daily absolute temperature maximum (55.5˚C) at Calarasi (Romania), and a minimum (-25.4˚C) at Izmail (south west Russia). In the west, maximum daily temperature (46.6˚C) for the 1980-99 period is found for Cordoba (Spain), and the minimum daily western temperature for the same period (-23.1˚C) was recorded at Clermont-Ferrand (southern France). Across the whole basin and also for the 1980-99 period, the greatest number (83) of frost days (days below 0˚C) for a single winter season was recorded for Embrun, also in southern France, and the longest (12 day) extreme summer heatwave (consecutive days over the 90$^{th}$ percentile of temperature, see Section 3.4.1), for Poretta-terme (northern Italy). Alcuescar (Spain) recorded the greatest number (89) of consecutive summer dry days (days with less than 1mm of rainfall each, see Section 3.4.1) and the daily rainfall maximum (408mm) for the period occurred at Alghero (Italy). Lindh (1992) found that minimum annual precipitation occurs in the southern Mediterranean regions (100-400mm/yr) and analysis of the FIC dataset shows minima at San Javier (74mm, 1961), Alcantarilla (80mm, 1966) and Alicante (88mm, 1995), all in southern Spain. Lindh (1992) found that the greatest annual precipitation values (1000-1500mm/yr) occurred on the coasts of the Adriatic and the Ionian Sea. Recorded values in the FIC data set show the largest values of annual total rainfall at Vigo Peinador (Spain, 2844mm), Penhas Dourados (Portugal, 2662mm) and Pescara (Italy, 2492mm), all in the central latitudes of the Mediterranean basin.

However, climate conditions may be considered extreme before such absolute extreme values are reached, and events can be labelled 'extreme' due to a combination of both persistence and magnitude. In the following sections, a number of 'extreme' events are discussed to inform the selection of extremes studied in this thesis, and to introduce other relevant issues. To these ends, each type of climatic extreme is explored in turn (Section 2.4.2) together with the processes that may be responsible for them (Section 2.4.3). Also introduced, in Section 2.4.4, is a key issue in studying extremes of climate, and of this study in particular: Observed changes in their frequency and magnitude.

### 2.4.2 Studying Mediterranean extremes

## Heatwaves

High mean summer temperatures across the majority of the Mediterranean basin create conditions such that hotter than average temperatures are likely to have major effects. Short bursts of high temperature, far enough above the mean (between 5 and 15˚C) that local water resources, infrastructure and population (particularly the elderly) cannot cope, can be defined as intense 'heatwaves' (Kysely, 2002). Record temperatures (e.g. 46.6˚C in Cordoba, Spain, during 1995) (Kysely, 2002), and wild fires (e.g. 430,000 ha of land burnt in Spain, plus 31 casualties, in 1994) (Piñol *et al.,* 1997) may be associated with such events. Mediterranean heatwaves tend towards one of two categories, short and intense (3-5 days and +7˚C-15˚C), or longer lasting and less pronounced (>10 days and ~+5˚C) (Colacino and Conte, 1995). Colacino and Conte (1993) describe both as being accompanied by intense drought and often forest fire. They show that the former, short events can have large areas of influence, including 'the entire territory of Italy… Corsica, Malta, the Adriatic side of the former Yugoslavia, Albania, part of Greece and of North Africa', and that there have been 32 of these short heatwaves from 1950-1992. The longer events may influence areas as wide as the entirety of the western and central basin and have numbered 28 over the same period. One example of such a long lasting event is the 1983 episode that lasted 20 days (between 13$^{th}$ July and 2$^{nd}$ August) and has been associated with 450 deaths in Rome.

The summer 2003 heatwave provides a recent example of a long lasting event (mid May to mid August) with a wide area of influence. Across the majority of western Europe (several million square kilometres), temperatures rose dramatically and a startling number of fatalities occurred (Munich Re, 2004). Estimates put figures at 14,802 heat related deaths in France, 2,045 in the United Kingdom, and 2,099 in Portugal (Koppe *et al.*, 2004), other estimates provide for more than 20,000 deaths throughout Europe (Munich Re, 2004), and up to as many as 40,000 (WWF, 2005), the vast majority of the year's fatalities (globally) due to heat stress (Munich Re, 2004). The heatwave cost in excess of US$10bn in insured losses, largely due to agricultural failure (Munich Re, 2004). Schönweise *et al.* (2004) found that these effects were generally due to near-record breaking temperatures during the day, and record breaking temperatures at night. Over the period daytime temperatures in excess of 40˚C occurred throughout France, Germany, Northern Italy, Spain, and Switzerland, while UK temperatures followed at a maximum of 37.8˚C. Night time temperatures exceeded 23˚C in both France and Germany, reaching maxima of 25.5˚C and 27.6˚C respectively. This 'climatic surprise' was probably the warmest European summer since 1540, a prolonged drought produced by a deep anticyclone over the English Channel (Pfister *et al.*, 1999; Luterbacher *et al.*, 2004). Schar *et al.* (2004) show the 2003 heatwave to be consistent with projected summer behaviour at the end of the 21$^{st}$ century. Projections show that the conditions associated with the 2003 heatwave (e.g. lack of convective rainfall, the positive feedback between soil moisture depletion and summer temperatures) may occur more frequently in the future.

## Cold snaps

Cold snaps are bursts of low temperature far enough below the mean that they stress health (Auliciems and Frost, 1989), energy use (Nguyen, 1994), and, to a greater extent, transportation (see below). Excepting locations at significant altitude (such as Alpine Italy, or parts of southern France), the only section of the Mediterranean basin with daily temperatures consistently (weeks to months) low enough (below 0˚C) to experience cold snaps with any frequency is the eastern basin (See Chapter 3.4). In the northern Balkans, exposure to winds from eastern Europe and the Former Soviet Union (FSU) may lead to temperatures persisting around –15˚C to -20˚C through January (Furlan, 1977). Throughout the entirety of the northern Balkan peninsula mean January temperature minima are negative (Furlan, 1977; FIC station data, 1958-99), in contrast with the western (Wallen, 1970; FIC station data, 1958-99) or southern (FIC station data, 1958-99) basin, where such conditions are rare. Little in the way of the literature deals with outbreaks of cold weather either in the eastern basin or across the rest of the Mediterranean.

During the recent harsh winter of 2004, reported by the U.S. National Climatic Data Center (NCDC, 2004), a period of exceptionally cold weather with numerous outbreaks of snowfall occurred across the majority of the eastern Mediterranean. From the $9^{th}$-$12^{th}$ January in Turkey 10 fatalities occurred and nearly 2000 villages were cut off by snowfall. At the start of that period temperatures fell as low as –13˚C (in Ankara), far below the normal monthly minimum of –3.3˚C. Towards the end of January a further seven fatalities occurred in north-eastern Romania, where 60 settlements were isolated, losing electrical service as well as road access. During February, heavy snows fell in parts of the Lebanon and Jordan (61cm of accumulated snow and three fatalities), Syria, and southern Greece, closing the Athens International Airport ($13^{th}$ February). Ancillary effects included an avalanche in south-eastern Turkey and, due to the winds and high waves that can accompany cold snaps, the loss of two ships off of the Turkish coast (NCDC, 2004).

Similar events occurred across the entirety of the Mediterranean basin during December 2001 (NCDC, 2004). From the $11^{th}$-$20^{th}$ of December heavy snow fell across Catalonia (NE Spain), La Rioja, Castilla, Leon (all central Spain), Venice, and Corsica, in addition to Greece, Turkey, and Poland. Traffic across these regions was heavily impaired, rendering Catalonia isolated for two days, Corsican villages inaccessible, Italian, French and Spanish roads chaotic, and the closure of all northern Greek airports and north-western Greek schools. Temperatures of –10˚C were experienced throughout

Spain and northern Greece, and in addition to heavy snow, intense rainfall and high winds were experienced across eastern Europe.

Prior to the 2001 event, a Greek cold surge in 1987 (3-13 March) became the worst Balkan snowfall of the preceding 100 years (Lagouvardos *et al.*, 1998) having a dramatic effect on Greek infrastructure. An example of a 'Balkan Front' (Metaxas, 1978), the cold wave propagated quickly (speed estimated at around $9ms^{-1}$) through the Balkan mountains from North and Central Europe into Greece. Causing elevated surface pressures (10hPa rise in 6 hours on $4^{th}$ March, 1987), strong north easterly winds ($24ms^{-1}$ at Limnos island) and large volumes of snow (2m depth in some mountain areas, snow every day from 4-13 March at Athens and non-coastal Greek stations), the severity of the cold surge was strongly related to its speed of onset (Lagouvardos *et al.*, 1998). Although Balkan Front initiated surges normally possess lifetimes of 1-3 days, the 1987 event caused widespread disruption in northern Greece and traffic difficulties between the north and south for 10 days. The movement of a pressure low from Italy into Greece, combined with flow from the north created a large cold air mass over the Greek peninsula for the period, and it was the inability of this air mass to progress over the Balkan mountains due to insufficient energy (a phenomenon termed 'cold air damming') that has been associated with the cold surge persistence (Lagouvardos *et al.*, 1998). Lagouvardos *et al.*, (1998), give a full synoptic analysis of the 1987 event, including more detail on cold air damming.

## Floods

Floods are highly complex events, related to numerous factors including intensity and persistence of rainfall, land use, and topography. The Mediterranean basin is particularly susceptible to flash floods with intense precipitation, and steep permeable slopes covered by sparse vegetation and thin soil (Belmonte and Beltran, 2001). Such events are common in the western Mediterranean, where flash flooding is considered the 'most destructive natural hazard' in terms of direct economic damage (Delrieu, *et al.*, 2005). Flash floods may also cause landslides, which increase the potential for destruction, In 1998 such an event caused 147 fatalities near Naples (Sarno-Quindici, Italy) and 29 million USD in damages (Lilljequist and Ligtenberg, 2005; Em-dat data set, 2002). A similar event in Algeria (9-10 November, 2001), affected 14 villages

(300million USD in damages) and caused an estimated 900 fatalities (Guerrieri, 2002). Named the worst Algerian flood on record (Tripoli *et al.*, 2002), 100mm of rain fell in 6 hours and winds of 33m/s were recorded, both caused by a particularly strong low-level cyclone exhibiting tropical characteristics.

Flooding due to intense rainfall (as opposed to that due to prolonged rainfall) is particularly prevalent in regions that display large variation in topography (Section 2.2.2) such as Alpine Italy (Brunetti *et al.*, 2001a; Egozcue and Ramis, 2001). Piedmont (in Northern Italy), for example, was affected by extreme precipitation during both November 1994 and October 2000. The former event (Cassardo *et al.*, 2002) occurred over 3 days, yielded rainfall intensities of up to 200mm/day, caused 64 fatalities, and 9313.5 million USD in damages (Em-dat, 2006), while the latter (Pelosini *et al.*, 2001) exceeded 600mm over 4 days, caused 29 fatalities and 434 million USD in damages (Em-dat, 2006). Along with local Mediterranean evaporation, and outside moisture sources such as the remnants of Extratropical Storm Leslie, the eastern Atlantic, and the African inter tropical convergence zone (Turato *et al.*, 2004), both rainfall events were intensified by orographic uplift (Pelosini *et al.*, 2001; Cassardo *et al.*, 2002). Significant positive divergences from annual maximum rainfall values for Genoa (positioned between mountains and the sea) have occurred only in 1842, 1970, 1977, 1992 and 1993 and in each case have caused dramatic flooding, causing both property damage and fatalities (Russo *et al.*, 2000). In 1970, the worst of the events, costs rose to 268million USD and 37 people died (Russo *et al.*, 2000).

The more vulnerable regions of eastern Spain (such as Valencia and Catalonia) exist to the north east of the region's main southern mountain range (Egozcue and Ramis, 2001) and to the south of the Pyrenees (Llasat and Puigcerver, 1994). Catastrophic flash flooding due to intense rainfall and orographic effect occurs frequently in Catalonia, one recent episode of which ($10^{th}$ June, 2000) caused 5 fatalities and 66 million USD in damages (Llasat *et al.*, 2003). Most weather stations in Catalonia and Valencia have recorded daily rainfall totals of more than 200mm (Font Tullot, 1983) and around Sierra de Aracena, the Gibraltar Strait, eastern Andalucia, Murcia, Valencia, parts of Catalonia, the Pyrenees, and Mallorca return periods for 100mm daily rainfall occurence are less than 5 years (Romero *et al.*, 1998a). Spanish floods of note include the events of 1962 (Barcelona, September 25th, 250mm of rainfall in 2 hours, 441 dead, 374 missing, 80 million USD), 1973 (Murcia, Almeria and Granada, October 19th, about 300 dead, 400 million USD), 1987 (Valencia, 3rd November, over 1,000mm in 36 hours in Gandia, 1,283 million USD), and 1996 (Huesca, August 7th, 200mm in 3 hours, 87 dead) (Em-dat, 2006).

The south of France is also susceptible to flood events. From 12-13 November, 1999, 35 fatalities in Carcasonne resulted from more than 550mm of rainfall in 24 hours (620mm in 48hrs) caused by the advection of moisture into the region from the Mediterranean, and frontal and orographic uplift (Bechtold and Bazile, 2001). Particularly dramatic episodes of flooding also occurred in the south of France during 2003. From the 2nd to the 4th of December of that year, torrential rainfall caused flooding throughout the Rhône valley region. Within 24 hours 300mm of rain fell, an 'amount that normally takes several months to accumulate' (Munich Re, 2004). In the September of 2002, Gard (in the Languedoc Rousillon region) experienced a precipitation maximum of 687mm in 24 hours, a 40 year return period event (Huet, 2003), 24 fatalities occurred, and economic damages totalled 1.2 billion euros (Delrieu *et al.*, 2005). The reason such a high cost was associated with the event is a function of its spatial extent: More than 200mm of precipitation fell over 5500km$^2$ of the Gard department during the 8th-9th of September period, a 140 year return period event (Huet, 2003). For more discussion of the causes of this rainfall event, see Section 2.4.3.

## Droughts

Drought is a multifaceted phenomena and can be measured in agricultural, climatic, hydrological, or socio-economic terms. Beyond those droughts associated with the heatwaves described above, climatological literature detailing large scale European drought is scarce (Lloyd-Huges and Saunders, 2002). Very intense heatwaves (such as the European 2003 event detailed above) may be accompanied by drought conditions. During the June phase of the 2003 heatwave, rainfall was almost completely absent from central and western Europe (Levinson and Waple, 2004). Although the socio-economic impacts of drought will be discussed in more detail in later chapters, it has been found in some cases (for Nebraska, US) that simple measures (like the Standardized Precipitation Index) of meteorological drought possess similar levels of utility to more complex measures, such as the Palmer Drought Severity Index (Oladipio, 1985; Lloyd-Huges and Saunders, 2000). It is the former that are largely dealt with in this study, as recommended by Bordi *et al.* (2001) for the treatment of Italian drought. The most simple measure of drought used in Mediterranean literature is the length of a given dry spell, as can be found in Douguédroit (1980), Kutiel (1985), and Anagnostopoulou *et al.* (2003).

Competition of tourism and agriculture for water use had led to an increased sensitivity to drought across the Mediterranean (Palutikof and Holt, 2004), a region that is naturally predisposed towards long dry periods (Section 2.1). Mediterranean summers often qualify as intense drought periods by the definition of the British Rainfall Organisation, a 15 day period with no more than 0.25mm of precipitation on any one day (Estrela *et al.*, 2000). Although in some areas (e.g. San Sebastian, northern Spain), the longest period over the last half century without rain has been only 40 days, these conditions can last for much longer, e.g. up to 190 days in Malaga, (southern Spain) (Martín-Vide and Gomez, 1999). Under a definition whereby a dry period consists of a consecutive sequence of days for which no single day accrues more than 10mm of rainfall, Almeria in south-eastern Spain may remain ‘dry’ for up to 360 days (Martín-Vide and Gomez, 1999). One of the regions in Spain to have been the subject of a study concerning drought is Valencia, a region known for flooding in autumn. Drought (i.e. a period during which rainfall totals are 60% below the mean for 2 years or more, as defined by the WMO) conditions are both frequent and persistent in the Valencian region and multi-annual events have occurred from 1952-55, 1963-64, 1978-85, and 1993-1995 (Estrela *et al.*, 2000). The 1978-85 drought period was one of the longest droughts to occur in the last half century within the Mediterranean. In Murcia, another part of southeast Spain, season length dry periods also occur. During the summers of 1994 and 1998 records show no rainfall and only trace amounts, respectively. In 1995 autumn rainfall at Murcia totalled only 15mm (Goodess, 2000).

The 2004/2005 western European drought event (Chapter 1.1) reduced rainfall amounts for southern Iberia (between October 2004 and June 2005) by roughly 60% (Garcia-Herrera, 2006). For Lisbon, these conditions resulted in the driest event in the last 140 years (Garcia-Herrera, 2006). However, in Portugal, drought is frequent and rainfall during March (historically a precipitation peak in some Portuguese regions) has been declining since the 1960s (Corte-Real *et al.*, 2000). The circulation responsible for these dry periods (an enhanced Azores high pressure zone situated to the west of Iberia) has been shown to be quasi-stationary, explaining the persistence of drought conditions for months, rather than weeks (Corte-real *et al.*, 2000).

In the Eastern Mediterranean, extremely dry years have been experienced in Turkey from 1990-91 (Nicosia, 116mm) and in Greece (1976-1977, Thessaly; 1989-90, Athens, 148.9mm; 1993, Athens), although across the region no synchrony in timing has been observed, suggesting localised underlying mechanisms (Kutiel *et al.*, 1996a). The 1989-1990 Greek event showed a 43% decrease in average yearly precipitation (almost zero precipitation for the usually wet month of January) - seriously affecting livestock, crops and wildlife, with losses estimated at 1.5 billion USD (Karavitis, 1998). Shortly after, in 1993, another serious drought occurred, reducing inflow into Mornos reservoir to 37.5% of the annual average. This second drought event subsequently caused both power failures in Athens, and flood problems from 'normal' precipitation events after the drought had broken (Karavitis, 1998). Very long dry spells (around 180 consecutive dry days), such as these, have also occurred in the central and southern Aegean Sea regions (specifically Rhodes and the Cyclades) (Anagnostopoulou *et al.*, 2003).

### 2.4.3 Extremes as a result of local dynamics

As with mean rainfall, intense cyclogenesis is likely to be a strong driving mechanism behind changes in extreme precipitation (Kostopoulou, 2003) Changes in the behaviour of intense rainfall can vary (spatially) over short distances, such as from one coast of northern Italy to the other, much like the average distribution (Brunetti, 2001a), and for similar reasons. Alpert *et al.* (2002) links increases in extreme rainfall frequency with increasing frequency and persistence of cyclones over the Mediterranean. Jansà (1995) states that Mediterranean cyclones are very important in organising flow (focussing and triggering deep convection), and thus for heavy rainfall associated with flooding in Alpine Italy, as seen during the events of September 1993.

Mediterranean cyclogenesis gives rise to torrential rainfall on the Spanish east coast, including Valencia (Estrela *et al.*, 2000) and Serra *et al.* (1999) show that enhanced precipitation in Catalonia is linked to eastern surface circulation, also a facet of local depressions. This eastern flow component of heavy Spanish rainfall is further explored by Goodess (2000) who describes it as a function of both orographic uplift, and advection from the Gulf du Lyon/Genoan Gulf region of the Mediterranean Sea. In a similar fashion, extremely wet conditions (standardized precipitation greater than or equal to 1.5) in Greece (Thessaloniki and Athens) have been linked to easterly flow from the Aegean (due to cyclones) and intense precipitation in Jerusalem has been attributed to travelling depressions from the north-west (Kutiel *et al.*, 1996a).

Intense rainfall has been associated with factors other than cyclogenesis. Estrela *et al.* (2000) show that orographically induced thunderstorms, due to the Iberian thermal low, can produce large volumes of precipitation. The particularly wide September 2002 storm (Gard, southern France) was associated with a large mesoscale convective system that resulted from a combination of the Mediterranean heat reservoir effect (see Section 2.2.2) and an upper level cold air trough extending southward from the UK into Iberia (Delrieu *et al.*, 2005). The combination of these two factors can create a southerly, unstable, and moist air flow toward the southern coasts of north-western Mediterranean countries, that is then lifted and channelled by orography to produce convection, and may result in very large storm systems. Topography may retard the progress of a mesoscale convective system such that is persists in one region for hours, rather than days (Delrieu *et al.*, 2005).

On a global scale, Alpert *et al.* (2002) demonstrate links between years with high frequencies of torrential rainfall (128mm/day) across the Mediterranean and strong El Niño years. Alpert *et al.* (2002) also show that contributions to total annual rainfall from torrential events occurring within El Niño years are becoming greater (7% of total annual rainfall from torrential events in 1953, about 9% in 1965, 15% in 1982/3, about 16% in 1986/7), as consistent with both Price *et al.* (1998) and Kadioglu *et al.* (1999).

In some regions of the Mediterranean, extreme dry conditions are associated with travelling cyclones. In Greece, south westerlies (i.e. Saharan sourced cyclogenetic flow) suppress precipitation (Kutiel *et al.*, 1996a). Reduction of precipitation has been associated with an increase in the frequency and persistence of sub-tropical anticyclones (Piervitalli *et al*. 1997). For lower latitudes synoptic subtropical patterns exert a greater influence on climate than for higher latitudes, and (particularly) in summer these patterns create conditions that may produce longer droughts (Martín-Vide and Gomez, 1999). Atmospheric stability is induced in the middle and upper atmosphere by the subtropical anticyclone, and where its influence is felt (i.e., the southern coast of Spain and the southern basin in general) precipitation is suppressed (Martín-Vide and Gomez, 1999). Droughts across the southern Aegean Sea have been linked to stability over Greece caused by the persistence of the subtropical anticyclone at height (Anagnostopoulou *et al*., 2003). Further, high NAO values can be linked to frequent anticyclones and dry conditions over the basin (Easterling *et al*., 1997) through the suppression of flow during periods dominated by the Azores anticyclone (Palutikof and Holt, 2004). This relationship is reinforced by Trigo and DaCamara (2000), who link extremely dry (wet) Portuguese years to low (high) frequencies of cyclonic, westerly, or southwesterly circulation.

Colacino and Conte (1995) suggest that short Mediterranean heatwaves (as defined in Section 2.4.2), and thus possibly forest fires, may be caused by a northward summer displacement of the subtropical jet stream, and a resultant warming of 850hPa temperatures (possibly due to the influence of the African air-mass). A convergence to the right of the displaced flow may create downward adiabatic compression and thus heating. The short-period movements of the jet stream are reflected in the short but intense heatwave conditions that are produced. The longer lasting heatwaves (again see Section 2.4.2) are associated with a jet-stream induced sequence of pressure centres that squeeze an anticyclone between two lows and progresses from west to east (an 'omega' wave), drawing north African air up into the Mediterranean basin over an extended period (supported for Prague by Kysely, 2002). There is an apparent periodicity of omega wave induced heatwaves over the Mediterranean basin around 20 years, with an increasing trend in heatwave frequency that may be linked (Colacino and Conte, 1995) to the MO (see Section 2.3.2).

In summary, the greatest influences upon extreme events in the basin would appear to be:

- Cyclogenesis, which creates and strengthens rain-bearing low pressure systems that may later produce heavy rainfall (e.g. for Valencia, Catalonia, Thessaloniki, and Athens) and flooding (e.g. Italy, 1993; Algeria, 2001; Italy and Spain, 2003), or carry dry arid air that suppresses rainfall (e.g. for Greece).
- The influence of southern conditions and subtropical anticyclones that suppress rainfall and can create both short-lived, intense, heatwave conditions and longer periods of unusually high temperature (e.g. Rome, 1983).
- The Asian airmass channels flow from the north east, reinforcing unusually cold conditions in the eastern Mediterranean (e.g. Greece, 1987).
- Iberian thermal conditions that generate flash flood inducing thunderstorms within the interior of the peninsula. These may then be channelled by southerly or westerly flow to cause flooding in other areas (e.g. Valencia, Spain; Gard, France, 2002)
- Pressure oscillations that exist within the basin (such as the MO) that control the path of anticyclones, and therefore contribute to heatwave frequency (Colacino and Conte, 1995).
- And general westerly flow associated with the jet stream. Westerly flow can directly create intense rainfall along Atlantic facing coasts of the Mediterranean, and aids generation of both dry anticyclones and wet cyclones with effects as given above.

As documented above, these factors may also be inter-related through the behaviour of the Azores anticyclone, part of the North Atlantic Oscillation.

### 2.4.4 Trends in Mediterranean extremes

Over the last half century we can expect to have seen changes in global mean temperature (IPCC, 2001). These changes may be evidenced in the extremes of climate

(Palutikof and Holt, 2004) such that the occurrence of 'droughts, floods, and wind storm[s] can be expected to change'. In this section studies detailing shifts in Mediterranean climate (both precipitation and temperature) over the last 50 to 100 years are discussed.

Easterling *et al.*, (1997) have shown that a manifestation of global warming is a shrinking diurnal temperature range (DTR) and that the warming is attended by increasing daily temperature maxima, and a greater increase in minima (e.g. Beniston *et al.*, 1994; Karl *et al.*, 1993; Heino *et al.*, 1999). Analysis reveals that in Iberia and northern Italy maximum temperatures are increasing more rapidly that their respective minima (Easterling *et al.*, 1997; Brunetti *et al.*, 2000a). Although the northern hemisphere DTR decline has been shown to end in 1979 (Vose *et al.*, 2003), uneven decreasing frequencies of cold snaps and frost days (as seen in Heino *et al.*, 1999, Frich *et al.*, 2002) and increases in heatwaves and warm days (as in Plummer *et al.*, 1999, Manton *et al.*, 2001, Frich *et al.*, 2002) continue to occur across much of the globe.

In the northern Balkans (specifically in Bulgaria) an increase in DTR (small, but statistically significant) is consistent, in terms of magnitude, with that in Italy and Iberia (Brazdil *et al.*, 1996). However, the mechanism is different, as the Balkan DTR change is apparently a result of decreasing (rather than increasing) temperatures. Balkan minimum temperatures have decreased more (-0.40°C) than maximum temperatures (-0.08°C) from 1951-1990. This cooling trend is not isolated: A similar directional trend in temperature is apparent in parts of Greece and Turkey, where cold-spell days are increasing (Klein-Tank *et al.*, 2002). In all of the above examples of temperature change, and indeed globally, minimum temperatures are more susceptible to change than maxima (Karl *et al.*, 1991). Globally the direction of change in extremes has mirrored the mean warming trend, but in some regions within the Mediterranean there exists a spatially discontinuous cooling of mean temperatures (Maheras *et al.*, 1999a), and in the Balkans the extremes follow this trend instead. Klein-Tank and Können (2003) show that Europe (and the western coast of Greece) is generally experiencing a decline in the frequency of significant cold extremes. However, it has also been reported that there are four stations in the south east of Europe (i.e. the Balkans and Turkey) that display a significant cooling trend: more specifically, a large and significant (at the 0.05 level) increasing trend found for winter cold-spell days, defined as the number of days exceeding six to fall below the 10$^{th}$ percentile of temperature distribution over 1976-1999 (Klein-Tank *et al.*, 2002).

Increasing DTR is significantly negatively correlated with monthly precipitation totals, between –0.37 and -0.71 dependent on season and location (Brunetti *et al.*, 2000a). The link between DTR and precipitation is possibly a function of decreasing amounts of cloud cover (Brunetti *et al.*, 2000a; Piervitalli *et al.*, 1997). Brunetti *et al.* (2000b, 2001a) have shown that for north-eastern Italy an increasing proportion of annual rainfall has occurred as a result of intense events, while rainfall event frequency has decreased, particularly over the last few decades. The number of rain days is decreasing (-8.3 days / 100 years) more than precipitation totals (no significant trend), particularly in spring and autumn. This behaviour results in a positive trend in precipitation intensity, that biases daily precipitation classes towards the extreme and causes an attendant increase in flood risk (Brunetti *et al.*, 2000b). Italian precipitation totals (from 42 stations) greater than 128mm/day (signifying 'torrential' rainfall) have increased from 1% to 4-5% of total rainfall from 1951-1999 (Alpert *et al.*, 2002). In addition to this, the return period of Italian ((Russo *et al.*, 2000; Brunetti *et al.*, 2001b) extreme rainfall events (defined as exceeding the 99.9$^{th}$ percentile of the 1920-1998 distribution, and exceeding the mean plus two times the standard deviation, respectively) has become shorter in both. Russo *et al.* (2000) state that over the period 1975-1995, the likelihood of Genoan flooding has increased, and that the return period of flooding due to intense rainfall has declined considerably, from 1000 to 100 years.

Lloyd-Hughes and Saunders (2002) state that the trends in extreme drought (defined as a Standardized Precipitation Index, a standardized, normalized precipitation series, of –2 or less) over Europe for the 20$^{th}$ century are statistically insignificant but that the longest (in terms of mean length) droughts (typically of 40 months duration) in Europe occur in northwest France, northwest Russia, and of relevance to this study, Italy. Bordi and Sutera (2002) state that dry (defined as conditions where a Standardized Precipitation Index is less than –1 over a 24 month period) areas have increased in southern Italy over the last three decades, and that if the drying trend continues in a linear fashion that southern Italy will be 'considered… a dry climate in less than one hundred years'. Southern Italian annual rainfall has decreased by 156mm from 1923 to 2000, an effect that has 'strengthened over the last 30 years' and caused a shift in drought periods from severe to extreme (Piccarreta, 2003). However, only the winter part of this trend (the bulk of the change, at 133mm), and the monthly trend for June are statistically significant (Piccarreta, 2003), contrasting with the significant positive northern rainfall trend (Brunetti *et al.*, 2004) for winter. No statistically significant trend in annual rainfall has been found for Athens from the 1890s to 1985 (Katsoulis and Kambetzidis, 1988).

Spanish rainfall frequencies (over 182 stations) at either of the extreme ends of the distribution (Torrential, 128mm/day and Light, 0-4mm/day) have increased significantly, meaning greater frequencies of both extremely dry and wet days (Alpert *et al.*, 2002). Consecutive dry days have increased over the 1958-2000 period across the majority of Iberia (8.6% of stations are significant at the 95% level, out of 50.1% with an increasing trend), while the number of heavy rainfall events has declined (19% and 47.6% respectively) (Haylock and Goodess, 2004). Maximum Iberian trends are +0.718 days/year for consecutive dry days, and -0.182 days per year for number of heavy rainfall events (Haylock and Goodess, 2004), both of which have been linked to the NAO by cluster analysis (see Section 2.3.3 and Chapter 3). Further, the total annual amount of Spanish rainfall has declined by over 15% from 1951-1990 (Piervitalli *et al.*, 1997). The combination of these factors produces a regime of less frequent, more intense precipitation, much like that described above for Italy.

Maheras *et al.* (2004) show that Greek winter and autumn rainfall has been declining (significantly) from 1958-2000. In the eastern Mediterranean extremely wet years (normalised rainfall greater than or equal to 1.5) are slowly being replaced with extremely dry years (normalised rainfall greater than or equal to -1.5) (Kutiel *et al.*, 1996a) with the exception of Cyprus and northern Israel, which display neither positive nor negative trends (Alpert *et al.*, 2002). To date Kostopoulou (2003) provides the most comprehensive study concerning extremes of the eastern Mediterranean. Her work develops an automated classification scheme for the area (5-25E, 25-55N) and performs an in-depth study of the resultant patterns. Therein the following trends over a 43 year period (1958-2000) are revealed:

- Both minimum and maximum temperatures significantly increase, with Tmin displaying the larger increase. This produces a reduction in DTR across the Eastern Mediterranean.
- Both the number of frost days and heatwave frequency display a significant positive trend within the interior of the Balkan Peninsula.
- There exists a 'large positive and significant trend in the maximum number of consecutive dry days' in the southern coastal parts of the easternmost region of the basin (Balkan peninsula, western Turkey and Cyprus). In the main the easternmost region displays a reduction in extreme rainfall over time, and an increase in dry periods. By contrast almost the entirety of the Italian peninsula shows significant increases in measures of extreme rainfall.

Despite decreases in mean precipitation over the majority of the basin, frequencies and intensities of extreme rainfall have been rising significantly across the Mediterranean basin and in the Alps (Alpert *et al.*, 1990; Beniston *et al.*, 1994). The study by Alpert *et al.* (1990) utilises common severity categories to characterise extreme rainfall across the entire basin, and may not accurately capture regional behaviour, providing instead a generalised view of the Mediterranean. In using static thresholds, rather than levels dictated by local rainfall distribution, some regional detail may have been lost. In the Third Assessment Report (IPCC, 2001) conditions are outlined wherein an increasing proportion of annual rainfall has occurred as a result of intense events, while the frequency of rainfall events decreases. Increases in rainfall intensity and decreases in rain days (as seen in this Section) have made the Mediterranean region as a whole vulnerable to both drought and flood (IPCC, 2001), and future exacerbation of this changing precipitation regime may worsen the problem considerably.

## 2.5 Implications for this study

From the review of available literature presented above, a picture of the climatic behaviour of the Mediterranean basin can be constructed. Further, possible links between different elements of the Mediterranean environment and surrounding phenomena can be theorised. For the purpose of this study several 'null hypotheses' can be developed- ideas that illustrate situations that are believed to be false. By attempting to prove these null hypotheses, this study can explore the relationships they concern.

1.) That extreme events behave uniformly across the region.

2.) And that the tail ends of surface variable distributions display no trend over the last half century.

3.) That the Mediterranean surface climate regime has no relation to the movements of hemispheric scale air masses (such as the Asiatic high, and the North Atlantic system), or more localised effects (such as the Mediterranean Oscillation).

4.) That this extends to the tail ends of surface variable (i.e., temperature and precipitation) distributions.

These are the principal areas of research for this study, as can be seen from the literature, unified approaches to Mediterranean climate extremes are largely absent to date. There is a particularly distinct lack of work concerned with extreme event trends in the eastern Mediterranean (including the south Balkan peninsula, Turkey and Cyprus). Hypothetical relationships that might be pursued to disprove the above null hypotheses include, but are not limited to:

1.) A link between the North Atlantic system and Mediterranean cyclogenesis, mediated by the position of the Azores High and southerly subtropical anticyclones.

2.) A relationship between the Asiatic pressure system and conditions in the east of the Mediterranean basin mediated by changes in induced winds and local pressures.

3.) Congruent changes in both precipitation and temperature range that may be correlated due to changes in cloud cover, and pressure distribution, and that may react disproportionately to a changing mean.

4.) The existence of a Mediterranean Oscillation in addition to the above, and thus an enhanced differential between east and west basin pressure conditions.

**Table 2.1: Köppen climate types relevant to this study (Rudloff, 1981, after Köppen, 1931), where t is annual average temperature. The strictly Mediterranean Cs Climate is given in bold.**

| | |
|---|---|
| *'B' Climates -* | 'Dry' - arid regions where evaporation exceeds rainfall. Climate is dry if total rainfall:<br>< 20t+280 (if over 70% of rainfall is in warmest 6 months)<br>< 20t+140 (if 30-70% of rainfall is in warmest 6 months)<br>Adapted from Köppen by Trewartha (1968). |
| 'Bs' | 'Steppe' – summer dry season, rainfall is greater than half the above threshold. |
| 'Bw' | 'Desert' – winter dry season, rainfall is less than half the above threshold. |
| *'C' Climates -* | 'Warm temperate rainy' - average temperature of coldest month less than 18˚C and greater than -3˚C Average temperature of warmest month over 10˚C. |
| 'Cs' | **'Mediterranean' - summer dry season, at least three times as much rain in wettest month of winter as in driest month of summer, the latter having less than 30mm precipitation.** |
| 'Cf' | 'Humid sub-tropical' - wet year round, at least 30mm of precipitation in the driest month, difference between wettest month and driest month less than for Cs or for Cw (winter dry season). |
| *Subdivisions -* | |
| 'a' | 'Hot summer' - average temperatures of warmest month over 71.6˚F (22˚C). |
| 'b' | 'Cool summer' - average temperatures of warmest month below 71.6˚F (22˚C), at least 4 months above 50˚F (10˚C). |
| 'k' | 'Cool, dry climate' – average annual temperature below 18˚C (eg Bsk - mid latitude deserts). |
| 'H' | 'Highland' - altitude is greater than 2500m. |