insufficient data or trends vary across these regions. This, combined with issues with defining events, leads to the assessment that there is *medium confidence* that globally the length and frequency of warm spells, including heat waves, has increased since the middle of the 20th century although it is *likely* that heatwave frequency has increased during this period in large parts of Europe, Asia and Australia.

2.6.2 Extremes of the Hydrological Cycle

In Section 2.5 mean state changes in different aspects of the hydrological cycle are discussed. In this section we focus on the more extreme aspects of the cycle including extreme rainfall, severe local weather events like hail, flooding and droughts. Extreme events associated with tropical and extratropical storms are discussed in Sections 2.6.3 and 2.6.4 respectively.

2.6.2.1 Precipitation Extremes

AR4 concluded that substantial increases are found in heavy precipitation events. It was *likely* that annual heavy precipitation events had disproportionately increased compared to mean changes between 1951 and 2003 over many mid-latitude regions, even where there had been a reduction in annual total precipitation. Rare precipitation (such as the highest annual daily precipitation total) events were *likely* to have increased over regions with sufficient data since the late 19th century. SREX supported this view, as have subsequent analyses, but noted large spatial variability within and between regions (Table 3.2 of Seneviratne et al., 2012).

Given the diverse climates across the globe, it has been difficult to provide a universally valid definition of 'extreme precipitation'. However, Box 2.4 Table 1 indicates some of the common definitions that are used in the scientific literature. In general, statistical tests indicate changes in precipitation extremes are consistent with a wetter climate (Section 7.6.5), although with a less spatially coherent pattern of change than temperature, in that there are large areas that show increasing trends and large areas that show decreasing trends and a lower level of statistical significance than for temperature change (Alexander et al., 2006; Donat et al., 2013a, 2013c). Using R95p and SDII indices (Box 2.4), Figures 2.33a and 2.33b show these areas for heavy precipitation amounts and precipitation intensity where sufficient data are available in the HadEX2 data set (Donat et al., 2013c) although there are more areas showing significant increases than decreases. Although changes in large-scale circulation patterns have a substantial influence on precipitation extremes globally (Alexander et al., 2009; Kenyon and Hegerl, 2010), Westra et al. (2013) showed, using *in situ* data over land, that trends in the wettest day of the year indicate more increases than would be expected by chance. Over the tropical oceans satellite measurements show an increase in the frequency of the heaviest rainfall during warmer (El Niño) years (Allan and Soden, 2008).

Regional trends in precipitation extremes since the middle of the 20th century are varied (Table 2.13). In most continents *confidence* in trends is not higher than *medium* except in North America and Europe where there have been *likely* increases in either the frequency or intensity of heavy precipitation. This assessment increases to *very likely* for central North America. For North America it is also *likely* that increases have occurred during the whole of the 20th century (Pryor et al., 2009; Donat et al., 2013c; Villarini et al., 2013). For South America the most recent integrative studies indicate heavy rain events are increasing in frequency and intensity over the contient as a whole (Donat et al., 2013c; Skansi et al., 2013). For Europe and the Mediterranean, the assessment masks some regional and seasonal variation. For example, much of the increase reported in Table 2.13 is found in winter although with decreasing trends in some other regions such as northern Italy, Poland and some Mediterranean coastal sites (Pavan et al., 2008; Lupikasza, 2010; Toreti et al., 2010). There are mixed regional trends across Asia and Oceania but with some indication that increases are being observed in more regions than decreases while recent studies focused on Africa, in general, have not found significant trends in extreme precipitation (see Chapter 14 for more on regional variations and trends).

The above studies generally use indices which reflect 'moderate' extremes, for example, events occurring as often as 5% or 10% of the time (Box 2.4). Only a few regions have sufficient data to assess trends in rarer precipitation events reliably, for example, events occurring on average once in several decades. Using Extreme Value Theory, DeGaetano (2009) showed a 20% reduction in the return period for extreme precipitation events over large parts of the contiguous USA from 1950 to 2007. For Europe from 1951 to 2010, Van den Besselaar et al. (2012) reported a median reduction in 5- to 20-year return periods of 21%, with a range between 2% and 58% depending on the subregion and season. This decrease in return times for rare extremes is qualitatively similar to the increase in moderate extremes for these regions reported above, and also consistent with earlier local results for the extreme tail of the distribution reported in AR4.

The aforementioned studies refer to daily precipitation extremes, although rainfall will often be limited to part of the day only. The literature on sub-daily scales is too limited for a global assessment although it is clear that analysis and framing of questions regarding sub-daily precipitation extremes is becoming more critical (Trenberth, 2011). Available regional studies have shown results that are even more complex than for daily precipitation and with variations in the spatial patterns of trends depending on event formulation and duration. However, regional studies show indications of more increasing than decreasing trends (Sen Roy, 2009; for India) (Sen Roy and Rouault, 2013; for South Africa) (Westra and Sisson, 2011; for Australia). Some studies present evidence of scaling of sub-daily precipitation with temperature that is outside that expected from the Clausius–Clapeyron relation (about 7% per degree Celsius) (Lenderink and Van Meijgaard, 2008; Haerter et al., 2010; Jones et al., 2010; Lenderink et al., 2011; Utsumi et al., 2011), but scaling beyond that expected from thermodynamic theories is controversial (Section 7.6.5).

In summary, further analyses continue to support the AR4 and SREX conclusions that it is *likely* that since 1951 there have been statistically significant increases in the number of heavy precipitation events (e.g., above the 95th percentile) in more regions than there have been statistically significant decreases, but there are strong regional and subregional variations in the trends. In particular, many regions present statistically non-significant or negative trends, and, where seasonal changes have been assessed, there are also variations between seasons (e.g., more consistent trends in winter than in summer in Europe). The overall most consistent trends towards heavier precipitation events are found in central North America (*very likely* increase) but assessment for Europe shows *likely* increases in more regions than decreases.

2.6.2.2 Floods

AR4 WGI Chapter 3 (Trenberth et al., 2007) did not assess changes in floods but AR4 WGII concluded that there was not a general global trend in the incidence of floods (Kundzewicz et al., 2007). SREX went further to suggest that there was low agreement and thus *low confidence* at the global scale regarding changes in the magnitude or frequency of floods or even the sign of changes.

AR5 WGII assesses floods in regional detail accounting for the fact that trends in floods are strongly influenced by changes in river management (see also Section 2.5.2). Although the most evident flood trends appear to be in northern high latitudes, where observed warming trends have been largest, in some regions no evidence of a trend in extreme flooding has been found, for example, over Russia based on daily river discharge (Shiklomanov et al., 2007). Other studies for Europe (Hannaford and Marsh, 2008; Renard et al., 2008; Petrow and Merz, 2009; Stahl et al., 2010) and Asia (Jiang et al., 2008; Delgado et al., 2010) show evidence for upward, downward or no trend in the magnitude and frequency of floods, so that there is currently no clear and widespread evidence for observed changes in flooding except for the earlier spring flow in snow-dominated regions (Seneviratne et al., 2012).

In summary, there continues to be a lack of evidence and thus *low confidence* regarding the sign of trend in the magnitude and/or frequency of floods on a global scale.

2.6.2.3 Droughts

AR4 concluded that droughts had become more common, especially in the tropics and sub-tropics since about 1970. SREX provided a comprehensive assessment of changes in observed droughts (Section 3.5.1 and Box 3.3 of SREX), updated the conclusions provided by AR4 and stated that the type of drought considered and the complexities in defining drought (Annex III: Glossary) can substantially affect the conclusions regarding trends on a global scale (Chapter 10). Based on evidence since AR4, SREX concluded that there were not enough direct observations of dryness to suggest *high confidence* in observed trends globally, although there was *medium confidence* that since the 1950s some regions of the world have experienced more intense and longer droughts. The differences between AR4 and SREX are due primarily to analyses post-AR4, differences in how both assessments considered drought and updated IPCC uncertainty guidance.

There are very few direct measurements of drought related variables, such as soil moisture (Robock et al., 2000), so drought proxies (e.g., PDSI, SPI, SPEI; Box 2.4) and hydrological drought proxies (e.g., Vidal et al., 2010; Dai, 2011b) are often used to assess drought. The chosen proxy (e.g., precipitation, evapotranspiration, soil moisture or streamflow) and time scale can strongly affect the ranking of drought events (Sheffield et al., 2009; Vidal et al., 2010). Analyses of these indirect indices come with substantial uncertainties. For example, PDSI may not be comparable across climate zones. A self-calibrating (sc-) PDSI can replace the fixed empirical constants in PDSI with values representative of the local climate (Wells et al., 2004). Furthermore, for studies using simulated soil moisture, the type of potential evapotranspiration model used can lead to significant differences in the estimation of the regions affected and the areal extent of drought (Sheffield et al., 2012), but the overall effect of a more physically realistic parameterisation is debated (van der Schrier et al., 2013).

Because drought is a complex variable and can at best be incompletely represented by commonly used drought indices, discrepancies in the interpretation of changes can result. For example, Sheffield and Wood (2008) found decreasing trends in the duration, intensity and severity of drought globally. Conversely, Dai (2011a,b) found a general global increase in drought, although with substantial regional variation and individual events dominating trend signatures in some regions (e.g., the 1970s prolonged Sahel drought and the 1930s drought in the USA and Canadian Prairies). Studies subsequent to these continue to provide somewhat different conclusions on trends in global droughts and/or dryness since the middle of the 20th century (Sheffield et al., 2012; Dai, 2013; Donat et al., 2013c; van der Schrier et al., 2013).

Van der Schrier et al. (2013), using monthly sc-PDSI, found no strong case either for notable drying or moisture increase on a global scale over the periods 1901–2009 or 1950–2009, and this largely agrees with the results of Sheffield et al. (2012) over the latter period. A comparison between the sc-PDSI calculated by van der Schrier et al. (2013) and that of Dai (2011a) shows that the dominant mode of variability is very similar, with a temporal evolution suggesting a trend toward drying. However, the same analysis for the 1950–2009 period shows an initial increase in drying in the Van der Schrier et al. data set, followed by a decrease from the mid-1980s onwards, while the Dai data show a continuing increase until 2000. The difference in trends between the sc-PDSI data set of Van der Schrier et al. and Dai appears to be due to the different calibration periods used, the shorter 1950–1979 period in the latter study resulting in higher index values from 1980 onwards, although the associated spatial patterns are similar. In addition, the observed precipitation forcing data set differs between studies, with van der Schrier et al. (2013) and Sheffield et al. (2012) using CRU TS 3.10.01 (updated from Mitchell and Jones, 2005). This data set uses fewer stations and has been wetter than some other precipitation products in the last couple of decades (Figure 2.29, Table 2.9), although the best data set to use is still an open question. Despite this, a measure of sc-PDSI with potential evapotranspiration estimated using the Penman–Montieth equation shows an increase in the percentage of land area in drought since 1950 (Sheffield et al., 2012; Dai, 2013), while van der Schrier et al. (2013) also finds a slight increase in the percentage of land area in severe drought using the same measure. This is qualitatively consistent with the trends in surface soil moisture found for the shorter period 1988–2010 by Dorigo et al. (2012) using a new multi-satellite data set and changes in observed streamflow (Dai, 2011b). However all these studies draw somewhat different conclusions and the compelling arguments both for (Dai, 2011b, 2013) and against (Sheffield et al., 2012; van der Schrier et al., 2013) a significant increase in the land area experiencing drought has hampered global assessment.