Modelling and forecasting northeasterly cold surges during the East Asian winter monsoon

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**Modelling and Forecasting Northeasterly Cold Surges during the
East Asian Winter Monsoon**

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**Interdisciplinary Graduate School
Institute of Catastrophe Risk Management**

2019

Modelling and Forecasting Northeasterly Cold Surges during the East Asian Winter Monsoon

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Institute of Catastrophe Risk Management

A thesis submitted to the Nanyang Technological University in partial
fulfilment of the requirement for the degree of

Doctor of Philosophy

2019

Statement of Originality

I hereby certify that the work embodied in this thesis is the result of original research, is free of plagiarised materials, and has not been submitted for a higher degree to any other University or Institution.

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18th March 2019

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Associate Professor Lo Yat-Man, Edmond

Authorship Attribution Statement

This thesis contains material from one paper published in the following peer-reviewed journal and from two papers accepted at conferences in which I am listed as an author.

Results from Chapter 4 is published as Kumar A, Lo EY, Switzer AD. High Resolution Numerical Modelling of Cold Surges. Symposium on now casting and very short range forecast. World Meteorological Organization. 2016. Hong Kong. <https://public.wmo.int/en/media/news/symposium-nowcasting-and-very-short-range-forecast>.

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The contributions of the co-authors are as follows:

Kumar A., Lo EY., and Switzer AD designed the problem and solution approach.

All authors contributed to the writing of the manuscript.

A handwritten signature in black ink that reads "Anupam Kumar". The signature is written in a cursive style with a large initial 'A' and 'K'.

18th March 2019

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Kumar Anupam

Abstract

Strong Cold Surges (CSs) commonly occur in East Asia during the North-East Monsoon and can severely affect human life and cause considerable economic losses. We analyzed three strong CS events (CSEs) that occurred over South China during the winters of 2008, 2009 and 2016 on two broad aspects. First, a regional modelling approach that combines high resolution regional climate modeling using the Weather Research and Forecasting (WRF) model together with datasets from reanalysis, satellite, and observational weather stations is used to model the CS in South China. Second, the influence of large scale synoptic meteorology on the outbreak of CS in East Asia is investigated using data analysis of reanalysis data sets and weather charts.

The WRF regional modelling approach using a two-way nesting technique reproduced the short range (hourly to daily), sudden increase in mean sea level pressure, the steep drop in surface temperature, and increased wind speed patterns during the three CSEs as observed at the Hong Kong weather station. The investigation of the large atmospheric systems over East Asia demonstrates that such CSEs originate, intensify and continue as unusually long persisting extreme cold advection due to the interaction of the three major atmospheric systems, Siberian Mongolian High, Aleutian Low and Jet Stream during the North-East Monsoon. We thus propose a new theoretical framework to explain the mechanism that leads to the development and formation of CSEs in East Asia.

We propose that a CS in South China is firstly pressure-gradient driven between the Siberian-Mongolian High and the South China Sea and is further influenced by the presence of an intense low-pressure systems developed near the coast of Japan. The latter also leads to the bifurcation of the CS air mass with the low-pressure system highly influencing the CS intensity and subsequent trajectory towards the South China Sea. We also propose new criteria based on daily and monthly pressure gradient for identifying and forecasting CS at time scales varying in days (i.e. short range) to a month (extended range) over East Asia. The methodology and criteria developed are also validated against past reported events. This proposed mechanism as based on the interaction of the three large atmospheric systems has significant potential to be used as a reliable forecasting tool. Such a tool can further provide advance information to decision makers and policy makers in developing response to weather extremes.

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Table of Contents

Chapter 1	Introduction.....	1
1.1	Motivation & Research Objectives	1
1.2	Outline of the thesis	9
Chapter 2	Literature Review	10
2.1	General Circulation Patterns of the Atmosphere	10
2.2	Monsoon Season and Weather Systems	16
2.2.1	General Synoptic Pattern of North East Monsoon	16
2.2.2	Weather System of East Asia during NEM	19
2.3	Modelling Extreme Weather Events	22
2.3.1	Theory of Numerical Models	24
2.3.2	Global and Regional Models	26
2.3.3	Dynamical & Statistical Downscaling	29
2.4	Cold Surges in South China	31
Chapter 3	Model and Data Sets	34
3.1	Weather Research and Forecasting Model	34
3.1.1	Introduction to Weather Research and Forecasting Model	34
3.1.2	WRF Physics	41

3.2	WRF Forcing Data & Data Sets	48
3.2.1	Reanalysis	50
3.2.1.1	NCEP FNL Data Sets	51
3.2.1.2	ERA 40 Data Sets	52
3.2.1.3	ERA Interim Data Sets	52
3.3	Satellite Data Sets	53
3.4	Observational Data Sets	53
 Chapter 4 Surge Analysis and WRF Modeling for South China and Hong Kong		55
4.1	Introduction to CSEs of 2008, 2009, and 2016.....	55
4.2	Model Configuration & Experimental Design for South China and Hong Kong ..	57
4.3	Model Evaluation & Case Studies	59
4.3.1	Case Study: CSE 24 th Jan 2016.....	62
4.3.2	Case Study: CSE & Cold Spell Event 24 th Jan -16 th Feb 2008.....	71
4.3.3	Case Study: CSE & Cold Front Event 15 th Dec 2009	77
4.4	Design of new operational criteria for forecasting Cold Surges Over South China and HK	80
4.5	Summary	90

Chapter 5	Large Scale Pattern of CS over East Asia	91
5.1	Jetstream	91
5.2	Siberian Mongolian High and Aleutian Low Systems	92
5.3	Possible Mechanism for the outbreak of CS	95
5.4	Identification of a unique pattern related to the onset of CS	101
5.4.1	Role of SMH & AL Systems	101
5.4.2	Role of JS & Jet Streak	112
5.4.3	Formation of LPS near Coast of Japan	123
5.5	Discussion & Summary	130
Chapter 6	Conclusions	149
References		152
Appendix A		175
Appendix B		178
Appendix C		181
Appendix D		184
Appendix E		195

List of Figures

Figure 1.1	Topography of regions in East Asia and Southeast Asia plotted from USGS data at 5m resolution.	4
Figure 1.2	Projected (a) mean surface temperature and (b) sea level rise by the end of 21st Century	5
Figure 2.1	(a) General circulation of Earth's Atmosphere (b) North hemisphere cross section showing JS and tropopause elevations	12
Figure 2.2	(a) Semi permanent pressure systems in July (b) Semi permanent pressure systems in January showing mean pressure systems and wind vectors at the surface	13
Figure 2.3	Schematic of the typical locations of the Jet Stream in the Northern Hemisphere at 200 hPa (Image from NASA)	15
Figure 2.4	General synoptic pattern for NEM	18
Figure 2.5	Map of SEA with Country boundaries	22
Figure 2.6	Global & Regional Model domains	28
Figure 3.1	WRF (a) Software Infrastructure (b) Modelling System Flow Chart	36
Figure 3.2	The ARW model terrain following hydrostatic vertical coordinate	37
Figure 3.3	Schematic depicting the different interactions between microphysics scheme with (a) only warm cloud process and (b) mixed cloud process	42
Figure 3.4	WRF physics interactions. Model physics parameterization schemes valid for South China, HK.	45
Figure 4.1	WRF model domain D1 and its nested domain D2	59
Figure 4.2	(a) Daily mean MSLP and (b) Min surface temperature comparison between HKO Station and corresponding WRF simulations at 00 UTC of each day	61
Figure 4.3	HKO Weather Chart for (a) 2 am on 24th Jan and (b) 2 pm on 24th Jan 2016	63
Figure 4.4	Daily (a) mean MSLP, (b) min temperature, and (c) mean wind speed from HKO Station over 1-31st Jan 2016	64

Figure 4.5	ECMWF, ERA Interim plot for MSLP ((a) & (b)), 10m Wind ((c) & (d)) and 2m temperature ((e) & (f)) on 18 UTC 23rd Jan 2016 and 06 UTC 24th Jan 2016 ...	66
Figure 4.6	ASCAT data for (a) 10 m wind speed and (b) 10 m wind direction for 12 UTC on 24th Jan 2016	67
Figure 4.7	HKO Station hourly plots of (a) MSLP and (b) temperature from 02 am to 02 pm local HK time on 24 th Jan 2016 (indicated by black box). WRF-D2 time series plots for (c) MSLP, (d) 2m temperature and (e) 10m wind from 2 am 23 rd Jan to 2 pm 24 th Jan 2016 (horizontal axis indicates hour from 02 am on 24 th Jan to 02 pm local HK time 24 th Jan 2016). WRF-D2 spatial plots for HK region (f) MSLP (g) 2 m temp at 9 am (01 UTC) on 24 th Jan 2016. WRF-D1 spatial plots for South China (h) 10 m wind & SST contour at 9 am on 24 th Jan 2016. In Figure f and g, HK region is represented in red colour while in Figure h, HK is represented by a blue dot and isotherms are represented by black lines. In Figures f, g, and h the coastal boundaries are represented by black thick line	70
Figure 4.8	ECMWF, ERA Interim plot for MSLP for (a) 00 UTC 24th Jan 2008 and (b) 00 UTC 16th Feb 2008.....	72
Figure 4.9	Daily (a) mean MSLP, (b) min temperature, (c) mean wind speed and (d) total rainfall from HKO Station over 24th Jan-16th 2008	73
Figure 4.10	WRF-D1 (a) MSLP and (b) temperature from 24th Jan to 16 th Feb 2008; (c) WRF 10m wind at 8 am on 25th Jan; (d) HKO Weather chart for 2 pm on 2nd Feb 2008; (e) WRF 10m wind at 8 am on 3rd Feb 2008	76
Figure 4.11	HKO Weather chart for (a) 00 UTC on 15 th Dec 2009 and (b) 00 UTC on 16th Dec 2009	78
Figure 4.12	(a) HKO Waglan station MSLP from 02 pm to 11 pm HK local time on 15th Dec 2009 (indicated by black box); WRF-D2 (b) MSLP; (c) 2m temperature;(d) 10m wind valid from 2pm on 15th Dec 2009 (horizontal axis indicates hour from 02 pm on 15th Dec to 00 am on 16th Dec local HK time); (e) WRF-D1 10 m wind for 13 UTC on 15th Dec 2009	79
Figure 4.13	ECMWF, ERA Interim plot for max MSLP (a) Jan 2016 (b) Dec 2009 (c) Jan 2008 and (c) Feb 2008 computed from ECMWF, ERA Interim based on the analysis field at 00 UTC, 06 UTC, 12 UTC, and 18 UTC	82
Figure 4.14	ECMWF, ERA Interim plot for MSLP (a) on 13 th Jan 2008 at 00 UTC; (b) on 24 th Jan 2008 at 00 UTC; (c) on 29 th Jan 2008 at 00 UTC; (d) on 3 rd Feb 2008 at 00 UTC; (e) on 7 th Feb 2008 at 00 UTC; and (f) on 13 th Feb 2008 at 00 UTC shows formation of frequent LPS during Jan-Feb 2008 between 140 °E - 180 °E and 30 °N - 55 °N	86

Figure 4.15	ECMWF, ERA Interim plot for 925 hPa Wind (a) on 24th Jan 2016 at 06 UTC; (b) on 15th Dec 2009 at 12 UTC; and (c) on 5th Feb 2008 at 12 UTC shows formation of low pressure system within the domain 140 °E - 180 °E and 30 °N - 55 °N	88
Figure 4.16	WRF hourly 2m temperature for Nantau (triangle), HKO (square) and Dong Dang (circle) from 18 UTC on 23rd Jan to 00 UTC on 24th Jan 2016 (horizontal axis indicates hour from 18 UTC on 23rd Jan to 00 UTC on 24th Jan 2016.....	89
Figure 4.17	ECMWF six hourly 2m temperature for Jinan, Fukuoka, Busan and Qingdao over 00 UTC on 23rd Jan to 00 UTC on 24th Jan 2016	89
Figure 5.1	Schematic of EAWM progressing towards south China, HK	93
Figure 5.2	Semi-permanent pressure systems during Jan	95
Figure 5.3	Monthly averaged SMH MSLP (a) Jan 2016 (b) Dec 2009 (c) Jan 2008 and (d) Feb 2008 computed from ECMWF, ERA Interim analysis field based on daily maximum values over 00 UTC, 06 UTC, 12 UTC, and 18 UTC. The coastal boundaries of Lake Baikal are represented by black thick line. The domain indicates the defined SMH.	98
Figure 5.4	Monthly averaged AL MSLP (a) Jan 2016 (b) Dec 2009 (c) Jan 2008 and (d) Feb 2008 computed from ECMWF, ERA Interim analysis field based on daily min values over 00 UTC, 06 UTC, 12 UTC, and 18 UTC. Coastal boundaries are represented by black thick line. The domain indicates the defined AL	99
Figure 5.5	Monthly mean wind plots for JS (a) Jan 2016 (b) Dec 2009 (c) Jan 2008 and (c) Feb 2008 computed from ECMWF, ERA Interim based on daily values at 00 UTC, 06 UTC, 12 UTC, and 18 UTC. The coastal boundaries are represented by black thick line	100
Figure 5.6	MSLP (hPa) over SMH domain for (a) Jan 2016; (b) Dec 2009; (c) Jan 2008; (d) Feb 2008 over the entire month, computed from ECMWF, ERA Interim at 00, 06, 12,18 UTC. Dotted lines in (a) to (d) indicates fitted linear trends	103
Figure 5.7	MSLP (hPa) over AL for (a) Jan 2016; (b) Dec 2009; (c) Jan 2008; (d) Feb 2008, computed from ECMWF, ERA Interim at 00, 06, 12,18 UTC. Dotted lines in (a) to (d) indicates fitted linear trends	105
Figure 5.8	PD (hPa) for (a) Jan 2016; (b) Dec 2009; (c) Jan 2008; (d) Feb 2008 between the SMH domain and the AL domain over the entire month, computed from ECMWF, ERA Interim at 00, 06, 12,18 UTC. Dotted lines in (a) to (d) indicates fitted linear trends	107
Figure 5.9	ECMWF, ERA Interim 925 hPa winds (m/s) for (a) 12 UTC on 18th Jan 2016; (b) 18 UTC on 22nd Jan 2016; (c) 18 UTC on 12th Dec 2009; (d) 18 UTC on 17th	

Dec 2009 (e) 06 UTC on 8th Jan 2008; (f) 00 UTC on 12th Jan 2008; (g) 06 UTC on 16th Feb 2008; (h) 00 UTC on 18th Feb 2008.....110

Figure 5.10 ECMWF, ERA Interim 925 hPa winds (m/s) for (a) 00 UTC on 26th Jan 2016; (b) 00 UTC on 21st Dec 2016; (c) 00 UTC on 17th Jan 2008; (d) 00 UTC on 21st Feb 2008111

Figure 5.11 Six hourly variation of JS (m/s) in (a), (c), (e) and AL (hPa) in (b), (d), (e) computed from ECMWF, ERA Interim valid from (a) & (b) 1st Jan-31st Jan 2016; (c) & (d) 1st Dec-31st Dec 2009; and (e) & (f) 1st Jan-31st Jan 2008115

Figure 5.12 Vertical Winds between 1000 hPa and 250 hPa, computed from ECMWF, ERA Interim for the month of (a) Jan 2016 at Lat 32⁰N; (b) Dec 2009 at Lat 31⁰N; and (c) Jan 2008 at Lat 32⁰N116

Figure 5.13 Jet Streak within JS (m/s) at 250 hPa, computed from ECMWF, ERA Interim valid for (a) 00 UTC on 19th Jan 2016; (b) 18 UTC on 13th Dec 2009; and (c) 4th Jan 2008117

Figure 5.14 Jet Streak within JS (m/s) at 250 hPa, computed from ECMWF, ERA Interim valid for (a) 18 UTC on 27th Jan 2016; (b) 19th Dec 2009 at 18 UTC; and (c) 9th Jan 2008 at 06 UTC118

Figure 5.15 Vertical Winds between 1000 hPa and 250 hPa, computed from ECMWF, ERA Interim valid for (a) 18 UTC on 27th Jan 2016 at Lat 37⁰ N; (b) 18 UTC on 19th Dec 2009 at Lat 30⁰ N; and (c) 06 UTC on 9th Jan 2008 at Lat 33⁰ N119

Figure 5.16 ECMWF, ERA Interim plot of PJ over East Asia towards Northern Pacific at 250 hPa valid on (a) 24th Jan 2016 at 00 UTC; (b) 24th Jan 2016 at 12 UTC; (c) 15th Dec 2009 at 12 UTC; (d) 16th Dec 2009 at 00 UTC; (e) 26th Jan 2008 at 00 UTC; (f) 15th Feb 2008 at 12 UTC; and (g) 15th Feb 2008 at 18 UTC123

Figure 5.17 (a) JS at 250 hPa; (b) Vertical Winds at Lat 36⁰N between 1000 hPa and 250 hPa; (c) MSLP chart from ECMWF, ERA Interim for 00 UTC on 24th Jan 2016 ...126

Figure 5.18 (a) JS at 250 hPa; (b) Vertical Winds at Lat 34⁰N and 35⁰N between 1000 hPa and 250 hPa; (c) Vertical Winds at Lat 35⁰N between 1000 hPa and 250 hPa; (c) MSLP chart computed from ECMWF, ERA Interim valid for 00 UTC on 21st Dec 2009127

Figure 5.19 (a) JS at 250 hPa; (b) Vertical Winds at Lat 36⁰N between 1000 hPa and 250 hPa; (c) MSLP chart computed from ECMWF, ERA Interim valid for 06 UTC on 17th Jan 2008128

Figure 5.20 (a) JS at 250 hPa; (b) Vertical Winds at Lat 36⁰N between 1000 hPa and 250 hPa; (c) MSLP chart computed from ECMWF, ERA Interim valid for 00UTC on 13th Feb 2008129

Figure 5.21 MSLP (hPa) over SMH and AL (a), (c), (e) and PD (hPa) between SH and AL (b), (d), (f) for the month of Jan, Dec and Feb computed from ECMWF ERA Interim based on 38 years of daily data from 1979-2017	132
Figure 5.22 Monthly averaged PD (hPa) computed from ECMWF, ERA Interim for the period 1979-2016	134
Figure 5.23 (a) Jet Streak embedded within the JS at 250 hPa; (b) Vertical Winds between 1000-250 hPa along the center of the Jet Streak at 32° lat; and (c) Vertical winds along line AB between 1000-250 hPa at the ‘entrance’ of Jet Streak; and (d) Vertical winds along line CD between 1000-250 hPa at the ‘exit’ of the Jet Streak valid for the month of Jan as averaged over 1979-2017, computed from ECMWF, ERA interim.....	138
Figure 5.24 (a) Jet Streak embedded within the JS at 250 hPa; (b) Vertical Winds between 1000-250 hPa along the center of the Jet Streak at 32° lat; and (c) Vertical winds along line AB between 1000-250 hPa at the ‘entrance’ of Jet Streak; and (d) Vertical winds along line CD between 1000-250 hPa at the ‘exit’ of the Jet Streak valid for the month of Feb as averaged over 1979-2017, computed from ECMWF, ERA interim	139
Figure 5.25 (a) Jet Streak embedded within the JS at 250 hPa; (b) Vertical Winds between 1000-250 hPa along the center of the Jet Streak at 32° lat; and (c) Vertical winds along line AB between 1000-250 hPa at the ‘entrance’ of Jet Streak; and (d) Vertical winds along line CD between 1000-250 hPa at the ‘exit’ of the Jet Streak valid for the month of Dec as averaged over 1979-2017, computed from ECMWF, ERA interim	140
Figure 5.26 Climatology of MSLP for SH and AL systems for (a) December; (b) January; and (c) February for the period 1979-2017 as computed from ECMWF, ERA Interim	142
Figure 5.27 The progression of continental polar air masses from Arctic towards Siberia as evidenced through strengthening of MSLP. The cold air masses as it reaches Mongolia (pinkish color) bifurcates into two major tracks (indicated by black arrow). The progression of JS is indicated by shaded arrow. Red box in the figure indicates the region of active convection during Dec, Jan and Feb while the violet box indicates the region of severe convection during Jan	145
Figure A.1. WRF, three hourly D2-plot at HKO station (22°18’07’’N; 114°10’27’’E) from 15th Dec-16th Dec at 00 UTC for (a) for MSLP (hPa); (b) for 2m Temperature (°C); and (c) for 10m Wind (m/s) for MSLP. Simulations were performed using KF, BMJ, GD, No_CU and GF schemes	177
Figure B.1 HKO daily forecast for Temperature for CSE 2016	179
Figure B.2 HKO daily MSLP (in black) and WRF-D1 daily MSLP (in red) from 24th Jan to 16th Feb 2008 at 00 UTC	180

Figure C.1 (a) Daily PBLH during Cold Surge 2008, (b) Hourly PBLH during cold surge 2009 and (c) Hourly PBLH during Cold surge 2016 at HKO (Lat:22° 18'07", Lon: 114° 10'27"). Horizontal axis in (a) indicates days from over 25th Jan-16th Feb 2008 at 00 UTC. Horizontal axis in (b) indicates hours from 15th Dec 07 UTC to 18 UTC and (c) indicates hours from 23rd Jan 23 UTC 24th Jan 10 UTC.....183

Figure D.1 ECMWF, ERA Interim plot for MSLP (a) on 1st Jan 2016 at 00 UTC; (b) on 5th Jan 2016 at 00 UTC; (c) on 7th Jan 2016 at 00 UTC; (d) on 11th Jan 2016 at 00 UTC; (e) on 12th Jan 2016 at 00 UTC; (f) on 13th Jan 2016 at 00 UTC; (g) on 15th Jan 2016 at 00 UTC; (h) on 19th Jan 2016 at 00 UTC; (i) on 20th Jan 2016 at 00 UTC; (j) on 25th Jan 2016 at 00 UTC; and (k) on 31st Jan 2016 at 00 UTC shows formation of frequent LPS during Jan-Feb 2008 between 140 °E - 180 °E and 30 °N - 55 °N.186

Figure D.2 ECMWF, ERA Interim plot for MSLP (a) on 2nd Dec 2009 at 00 UTC; (b) on 4th Dec 2009 at 00 UTC; (c) on 7th Dec 2009 at 00 UTC; (d) on 13th Dec 2009 at 00 UTC; (e) on 14th Dec 2009 at 00 UTC; (f) on 16th Dec 2009 at 00 UTC; (g) on 17th Dec 2009 at 00 UTC; (h) on 18th Dec 2009 at 00 UTC; (i) on 19th Dec 2009 at 00 UTC; (j) on 20th Dec 2009 at 00 UTC; (k) on 21st Dec 2009 at 00 UTC; (l) on 22nd Dec 2009 at 00 UTC; (m) on 27th Dec 2009 at 00 UTC; (n) on 28th Dec 2009 at 00 UTC; (o) 29th Dec 2009 at 00 UTC; (p) on 30th Dec 2009 at 00 UTC; and (q) on 31st Dec 2009 at 00 UTC shows formation of frequent LPS during Jan-Feb 2008 between 140 °E - 180 °E and 30 °N - 55 °N.....191

Figure D.3 ECMWF, ERA Interim plot for MSLP (a) on 6th Jan 2008 at 00 UTC; (b) on 7th Jan 2008 at 00 UTC; (c) on 9th Jan 2008 at 00 UTC; (d) on 10th Jan 2008 at 00 UTC; (e) on 3rd Feb 2008 at 00 UTC; (f) on 4th Feb 2008 at 00 UTC; (g) on 7th Feb 2008 at 00 UTC; (h) on 8th Feb 2008 at 00 UTC; (i) on 10th Feb 2008 at 00 UTC; (j) on 11th Feb 2008 at 00 UTC; and (k) on 13th Feb 2008 at 00 UTC shows formation of frequent LPS during Jan-Feb 2008 between 140 °E - 180 °E and 30 °N - 55 °N.....194

List of Tables

Table 4.1	WRF skill scores of MSLP and surface temperature at daily level over HKO for the period from 1st Jan -31st Jan 2016.....	61
Table 4.2	Pressure Gradient between SMH and SCS, and between SMH and LPS from ECMWF, ERA Interim for HK cold surge events (non-shaded cells) and HK non-cold surge events (shaded cells).....	84
Table 4.3	Max MSLP Values from ECMWF, ERA Interim within the SMH domain (40 ⁰ N -60 ⁰ N to 80 ⁰ E-120 ⁰ E for Jan 2016, Dec 2009 and Jan-Feb-2008	85
Table 5.1	Monthly average SMH, AL and their PD computed over period 1979-2017	133
Table 5.2	Possible CSEs years for Dec, Jan and Feb based on monthly averaged PD from the period 1979-2017	135
Table 5.3	CSE scale criteria for classifying CSE and their progression based on monthly averaged PD (hPa) and Jet Streak (m/s) from 1979-2017	146
Table 5.4	Reported CSEs, countries affected, losses in USD available from EM-DAT. Monthly avg. PD, Pressure Gradient, intensification of SMH & AL system and shifting of AL system	148
Table B.1	Monthly mean values and Normal (1981-2010) for Jan 2016 for surface pressure, min and mean temperature, mean wind speed and total rainfall at HK.....	178
Table B.2	Monthly mean values and Normal (1971-2000) for Jan 2008 for surface pressure, min and mean temperature, mean wind speed and total rainfall at HK.....	178
Table B.3	Monthly mean values and Normal (1971-2000) for Feb 2008 for surface pressure, min and mean temperature, mean wind speed and total rainfall at HK.....	178
Table B.4	Monthly mean values and Normal (1971-2000) for Dec 2009 for surface pressure, min and mean temperature, mean wind speed and total rainfall at HK.....	179
Table E.1	Monthly mean values and Normal (1979-2016) for Jan 2016 for SMH, AL and JS.	195
Table E.2	Monthly mean values and Normal (1979-2016) for Jan 2008 for SMH, AL and JS.	197

Table E.3	Monthly mean values and Normal (1979-2016) for Feb 2008 for SMH, AL and JS.	199
Table E.4	Monthly mean values and Normal (1979-2016) for Dec 2009 for SMH, AL and JS.	201

Acronyms

AL	Aleutian Low
AR5	Fifth Assessment Report (of the IPCC)
AMS	Asian Monsoon System
ASCAT	Advanced Scatterometer Data Products
CS	Cold Surge
CSE	Cold Surge Event
EAM	East Asian Monsoon
EAWM	East Asian Winter Monsoon
ECMWF	European Centre for Medium Range Weather Forecast
FNL	Final
GCMs	Global Climate Models
HK	Hong Kong
HKO	Hong Kong Observatory
IPCC	Intergovernmental Panel on Climate Change
ITCZ	Inter Tropical Convergence Zone
JS	Jet Stream
LPS	Low Pressure System
MK	Mann-Kendall
MSLP	Mean Sea Level Pressure
NCEP	National Centers for Environmental Prediction
NEM	North East Monsoon
PJ	Polar Jet

RCMs	Regional Climate Models
SCS	South China Sea
SEA	South East Asia
SH	Siberian High
SJ	Sub-Tropical Jet
SMH	Siberia Mongolia High
SST	Sea Surface Temperature
USGS	United States Geological Survey
UTC	Coordinated Universal Time
WRF	Weather Research & Forecasting Model

Chapter 1

Introduction

1.1 Motivation & Research Objectives

One of the major concerns related to potential changes in climate is the increase in climatic extremes and particularly on their location, intensity and frequency. There have indeed been significant changes in past climatic extremes as based on observational studies (e.g., Easterling et al. 2000; Alexander et al. 2006; Bronnimann et al. 2008; Houghton 2009). To understand the future implication, Global Climate Model (GCM) outputs have been analyzed for projecting future climatic extremes wherein they show increases in extreme events under the projected future climate (Fischer and Schar, 2009; Kjellstrom et al., 2007; IPCC, 2007; Tebaldi et al., 2006). An authoritative source of information on past, present and future climate state on environment and socio-economic impact is provided by the Intergovernmental Panel on Climate Change (IPCC) set up in 1988 by United Nations Environment Programme and World Meteorological Organization.

The IPCC Fifth Assessment Report (AR5) (IPCC, 2014) shows that the global climate has undergone a significant warming. The most existential threat arises from sea level rise that has already impacted coastal regions and exacerbated flooding from powerful storms and typhoons. As noted in AR5 report “Climate change is perhaps the single greatest challenge confronting the Asia-Pacific region, and its more than 4 billion people”. It has been observed that the warming trends over the continental Asia are the strongest and in particular the period from 1979 onwards, the warming is strongest for northern and eastern Asia in autumn and spring and over China in winter (IPCC, 2007). “As a part of the polar amplification, large warming trends ($>2^{\circ}\text{C}$ per 50 years) in the second half of the 20th century are observed in the northern Asian sector (IPCC, 2013)”. The warming trend over the period 1901-2009 is particularly strong with an increase of 2.4°C during the cold

season between November and March in the semiarid area of Asia. A similar increasing annual mean temperature trend has been observed in South and East Asia during the 20th century (IPCC, 2014). Furthermore, Dong et al. (2010) noted that “different regions have different warming rates and sea level rise.” For South China and Hong Kong (HK), the annual mean sea level, relative to the average level of 1986-2005 is expected to rise by 0.63 - 1.07 m for the period 2081-2100 (HKO, 2017). For south China where its coastal regions receive severe threats from intense and frequent storms from the South China Sea (SCS) the changing climate as it impacts on the SCS needs to be considered.

Since temperature has globally risen, large regional variations and considerable warming trends have been reported for East Asian countries (e.g. Trenberth et al., 2007). Also, the various numerous models ranging from simple climate models to atmospheric and oceanic coupled global climate models have reported that almost everywhere cold days are projected to significantly decrease in the future with global warming (e.g. Meehl et al., 2007). All these possibly suggest that global surface warming inevitably reduces the number of cold days (e.g. Alexander et al. 2006). However, its effect on the frequency of cold extremes at regional scales still remains controversial (Park et al. 2011, Choi et al. 2009).

Sea level rise is considered as one of the important societal aspects of climate change, though there is a large uncertainty about the expected changes from it (Stammer and Hüttemann, 2008; Kuhlbrodt and Gregory, 2012). There is no direct link that has been established or reported between the sea level rise and Cold Surges (CSs) but the development of these cyclones have also been linked to the CSs (Takahashi et al. 2011). It has also been suggested that ‘extreme’ sea level is exacerbated by tropical cyclones which will lead to accelerated sea level rise (Woodruff et al., 2013).

The SCS is one of the biggest marginal seas in the world. As shown in Figure 1.1, the SCS lies in the western Pacific Ocean that borders the South East Asia (SEA) mainland and has a latitudinal extent between 10 °N and 20 °N and longitudinal extent between 110 °E to 120 °E. The near surface circulation of SCS is primarily driven by monsoon winds

(Hu et al., 2000; Wyrki, 1961). It further plays a major role in significantly influencing the regional climate in the coastal regions. In terms of climate changes, Figure 1.2 reproduced from AR5 report indicates that temperature over the East Asia region (encircled) is projected to increase by 3-5 °C (Figure 1.2a) under the various Representative Concentration Pathway (RCPs) scenarios. The corresponding sea level is projected to rise from 0.4-0.6 m (Figure 1.2b).

Figure 2.1: Topography of regions in East Asia and Southeast Asia plotted from USGS data at 5m resolution. In the figure South China Sea is marked as SCS.

Data Source: http://www2.mmm.ucar.edu/wrf/users/download/get_sources_wps_geog.html

Figure 1.2: Projected (a) mean surface temperature and (b) sea level rise by the end of 21st Century (Reproduced from IPCC, AR5 (2014)). East Asian regions are encircled in the figures.

As a part of the northern winter monsoon, the northeasterly trade winds from the equator to 20° N dominates the tropical region of the western Pacific and eastern Indian Ocean. The northerly flow undergoes periods of intensification over the extreme western Pacific and SCS that persists over synoptic time scales. During this period of North-East Monsoon (NEM) / winter monsoon, one of the most distinct weather events that occurs

over November to February are the CSs (Chen et al. 2002). East Asian CSs are typically associated with a sudden drop in temperature and a change of wind direction to northerly from easterly. They occur once or twice per month and may last from a few days to a week or even longer [Wu and Chan (1997), Zhang et al. (1997) and Wu and Chan (1995)]. This intensification of northeasterly flow leading to CSs in the mid-latitudes is due to the formation of migratory anticyclones over Siberia and Mongolia which will be referred to as the Siberian Mongolian High (SMH). The onset and occurrences of CS in East Asia are significantly correlated with the intensity of the SMH and its southeast propagation (Ding 1990; Wu and Chan 1997; Wang and Ding 2006).

The wintertime temperature over East Asia has a relatively high intra-seasonal variability (Gong and Ho 2004) and it is also one of the major regions under the significant influence of CS (Ding et al. 2009). Approximately 10 CSs sweep across East Asia each winter (Chen et al. 2004). The cold and dry continental air masses associated with the CS causes an intense drop in temperature over south/southeast Asia as it progresses southward. This leads to the freshening of the northeasterly flow in the SCS which can impose four major dramatic effects on convective patterns on synoptic and planetary-scales. Of these, the first effect contributes towards a general increase of convective activity with deep convection over maritime continent (Chang et al., 2011) during the NEM with strong CSs leading to heavy rainfall and are often associated with severe flooding in the equatorial zone particularly over Malay Peninsula, Sumatra, Borneo and localized flooding (Johnson and Chang 2007; Tangang et al. 2008). However, the strongest CSs are concentrated in the SCS where they can reach and cross the equator because of orientation of regional topography (Figure 1.1) which acts to restrict the flow such that the low level north easterly flow is channeled toward the equator (Chang et al. 2016). For instance, the CS that occurred from 10th January to 5th February 2008 induced extremely damaging ice storms, snow and frosts in south China (Yang et al. 2010; Lu et al. 2010) killed 129 people, damaged 11867 kilo/hectare crops and caused \$24 billion in economic losses (Zhao et al 2008; DCAS/NCC/CMA 2008). Secondly, since CSs are associated with sudden changes in pressure and temperature (Wu and Chan 1995; Chang et al. 1979), the

occurrences of CSs can have massive impacts both on the human activities and environment (Yang et al. 2009; Lin et al. 2009). Thirdly, the northeasterly CSs may have a significant impact on the development of tropical cyclones (Takahashi et al. 2011). Lastly, these surges, as they progress towards SEA may also have a possible influence on the monsoon system of the southern hemisphere and on the formation of the tropical cyclones therein.

The Earth's global climate is actively being modelled using GCMs. GCMs simulate the Earth's climate via mathematical equations that describe oceanic, atmospheric, biotic processes, feedback and interactions. Today they are considered as valuable predictive tools that provide reasonably accurate climate information for present and future climate on global and regional scale. However, due to the coarser resolutions of GCMs, they cannot account for fine-scale heterogeneity of climate variability. Since decision makers require information on potential impacts of weather extremes as such these heterogeneities are of high importance. Climate Change impact studies of a specific region using coarse resolution GCM (as compared to e.g. regional models) result in the smoothening of land surface representation in the analysis. These analyses yield unrealistic regional climate information even though the large-scale climate might be well represented. In contrast, Regional Climate Models (RCMs) have proven to provide better projections of future climate at a regional level (e.g., Lo et al. 2008; Wang et al. 2004). A process termed downscaling has been developed in order to derive climate projections at scales that decision makers desire. A large scientific research community today is successfully using RCMs to dynamically downscale the global climate information from GCMs to have more localized results. The process of downscaling captures sub-grid scale contrasts and inhomogeneities so that information is more realistic at a finer scale and thus adds information to the coarse GCM output when downscaled. Detailed information of such work and successful performance of RCMs are available e.g. in Feser (2011), Foley (2010), Rummukainen (2010), and Wang et al. (2004).

However, even RCM modelling of CS Events (CSEs) in South China and the SCS has always been a challenge due to its geographical location which has a high influence from

the surrounding air-sea interactions making the weather pattern highly complex. Furthermore, for the countries therein to adapt to climate change and in particular for the less developed countries, it is necessary to have estimates of the range of likely climatic responses or at least an idea of the most likely response for a given period in the future. So far, very few reported studies have focused on the occurrences of winter CSEs in southeast China. A recent study by (Park et al. 2011a) indicates that CSEs tended to occur less frequently over China but more frequently over Japan and Korea. This indicates that some cold air outbreaks in East Asia may not affect South China and thus the occurrences of CSEs in East Asia may not reflect the change in the occurrence of CSEs in South China. Moreover, despite an overall decreasing trend in number of cold days, the CSEs frequency in East Asia is reported to have been stable for the last few decades (Park et al. 2011b). Thus, it is of interest to better forecast future changes of CSEs by investigating the possible mechanism that leads to its occurrences. However, such an investigation will require a detailed understanding of the complex dynamic mechanisms of CS and associated atmospheric circulation patterns. This occurrence is also needed for a reasonable estimate of the occurrence and intensity of CSEs under the influence of changing climate.

The PhD thesis thus aims to model the distinct extreme CSEs and to investigate the possible mechanisms leading to a CSE outbreak. The goal is to provide new criteria as based on newly proposed theoretical framework for increasing the accuracy of CSE predictions in East Asia. The objectives of this thesis are first to improve our understanding of CSEs in South China through observations and regional model simulations at different spatial scales, and second to develop a new theoretical framework by incorporating large-scale atmospheric circulation patterns for predicting the occurrence of CSEs in East Asia and its subsequent progression towards SEA. For this we analyzed three strong CSEs that occurred over South China during the NEM period in 2008, 2009 and 2016. Firstly, we used a regional modelling approach that combines high resolution regional climate modeling together with datasets from reanalysis, satellite, and observational weather stations. Secondly, we investigated the influence of large scale

synoptic meteorology on the outbreak of CSEs in East Asia via data analysis of reanalysis data and weather charts.

1.2 Outline of the thesis

The structure of the PhD thesis is as follows:

1. *Chapter 1* provides an introduction and motivation of the PhD research work.
2. *Chapter 2* provides a literature review related to the general synoptic pattern of North East Monsoon and weather systems in East Asia. This chapter further provides information on (i) numerical modelling, (ii) global & regional climate models and (iii) past modeling of CSEs in South China.
3. *Chapter 3* describes the physics of the regional climate model, the Weather Research and Forecasting (WRF) model, the various data sets used for the analysis such as ERA Interim, NCEP-FNL, ASCAT, and HKO station data.
4. *Chapter 4* describes the WRF model configuration and experimental design for regional modelling of the CS in the South China and HK region. The results lead to a new criterion for forecasting these events in South China and HK.
5. *Chapter 5* describes analysis of the large atmospheric systems comprising the Siberian High, Aleutian Low, Jet Stream and their interactions. These interactions define a new theoretical framework for predicting the outbreak of CS in East Asia during the NEM months of Dec, Jan & Feb.
6. *Chapter 6* represents conclusions of the thesis work.

Chapter 2

Literature Review

2.1 General Circulation Patterns of the Atmosphere

The planetary or global wind system is considered the main transport mechanism for moisture and energy through the Earth's atmosphere. The planet Earth is heated unevenly by the incoming solar radiation. This heat is carried away from the hotter region (tropical) to the colder region (polar) by the large-scale geophysical atmospheric flows known as "general circulation". The atmospheric circulation and the ocean heat transport then act together to regulate Earth's climate through an exchange of latent heat and dry static energy (Trenberth and Stepaniak 2003a, b; Trenberth et al. 2001). These exchanges of energy are therefore important in the understanding of the global patterns of precipitation (Muller and O'Gorman (2011)). Surface air diverges from a high-pressure region and converges to a low-pressure region. Thus, to understand the global wind patterns, it is important to delineate two important characteristics of atmospheric circulation, 'convergence' and 'divergence' that determine where the air will be rising or sinking. The wind direction at the various levels of the atmosphere also drives the local climate and weather system of a region. In turn, the general direction of the winds is dependent on the differences in atmospheric pressure that include factors such as the latitude and its proximity to oceans. Figure 2.1a shows the general circulation pattern of the Earth's atmosphere. The Hadley cell is the largest of the cells in the atmosphere, extending from the equator, where the air starts rising and ends roughly at 30° latitudes where it starts sinking (Figure 2.1b). These cells are responsible for the occurrence of trade winds in tropics and control the weather patterns at these low latitude regions. The Ferrel cell generally occurs between the sinking air near 30° latitudes and rising air farther poleward at roughly 60° latitudes (Figure 2.1b). This circulation belt is also known as prevailing westerlies. The circulation within the Farrel cell is complicated as it has a sinking air

where it is relatively warmer and rising air where it is relatively colder. The Polar cells are the weakest and the smallest cells that extend from about 60° to the poles (Figure 2.1b). Within this cell, the cool air sinks at the poles and flows towards the lower latitude. In the tropics, the trade winds within the Hadley Cells are considered to be extremely important in determining the interactions of weather and oceans that affect the general circulation pattern. These winds have an average speed from 4 to 7 m/s and the speed further increases during the winter season. Figure 2.2 shows the mean pressure system and wind flow at the surface. The red dotted line in the figure represents the Equatorial Trough where the trade winds tend to converge. During the month of July, most of the prominent low-pressure centers (Figure 2.2a) are the thermal lows over North America (Sonora Low), over Africa (Sahara heat low) and over India. During the month of January (Figure 2.2b), most of the prominent low pressures centers are located over the subtropics of Southern Africa, Southern America and Australia Indonesia.

Figure 3.1: (a) General circulation of Earth's Atmosphere (b) North hemisphere cross section showing Polar Jet, Subtropical Jet and Tropopause elevations.

Source: <http://www.metoffice.gov.uk>; <https://www.weather.gov/jetstream/jet>.

Figure 2.2: Semi permanent surface pressure systems in (a) July (b) January showing mean pressure systems and wind vectors at the surface. The red dotted line in the figure represents the Equatorial Trough where the trade winds tend to converge.

Source: <http://www.meted.ucar.edu/>

In addition to the three-cell circulations as discussed above, another ubiquitous structural characteristic of Earth's atmosphere are the Jet Stream (JS). These are rapidly flowing, narrow currents of air situated near the tropopause between the troposphere and the stratosphere. The troposphere contains two principal JS (Figure 2.3), viz., the westerly polar JS (also known as polar front) of the winter and summer hemispheres; and westerly subtropical JS of winter and summer hemispheres & easterly subtropical JS of the summer hemisphere (Rossby et al., 1947). The appearance of JS results for the complex interaction between various atmospheric variables such as warm and cold air, high and low pressure systems and seasonal changes. They meander around the globe, rising and dipping in altitude/latitude, forming eddies and splitting at times and even disappearing altogether to appear somewhere else. The fastest JS are the Polar Front Jet, commonly known as the Polar Jet (PJ), and the Subtropical Jet (SJ) which meander in a wave-like pattern around the globe from west to east (Figure 2.3). The regions around 30° N and 30° S near 200 hPa are known for the location of SJ and regions around 50° to 60° N and 50° to 60° S between 300 and 200 hPa are known for location of PJ (Figure 2.1b). The PJ forms in the region of strong temperature gradients between warmer air masses and the cold, polar air and is strongest during winter when it migrates to tropical latitudes and it may merge with the SJ.

The first comprehensive descriptions of the JS structure over the north Pacific was provided by Mohri (1953). The study emphasized that the JS is located between contrasting air masses and this single entity JS is in fact a hybrid of the separate SJ and PJ. Numerous attention has been given towards understanding the influence of external processes on the evolution of the JS. For instance, on a synoptic scale, deep tropical convection can impact the JS either directly via upper-level divergent outflow (e.g., Archambault et al. 2013) or remotely through downstream baroclinic development (e.g., Riemer and Jones 2010; Higgins et al. 2000; Madden and Julian 1994; Kiladis and Weickmann 1992). Studies of JS observational characteristics have also been reported (Jensen 2015; Yang et al. 2002; Wang et al. 2014). The increasing number of mid-latitude weather extremes in recent decades have led to a wider research interest in investigating

the possible role of JS in modulating these extremes (Rikus 2015; Woollings et al. 2014). However, the recent interest in linking JS is not only limited to the meteorological and climate modeling community but has permeated into research fields that study the impact of climate and future climate change on ecosystem and societal dynamics (e.g., Stark et al. 2016).

Figure 2.3: Schematic of the typical locations of the Jet Stream in the Northern Hemisphere (Image from NASA).

Source: <https://climate.ncsu.edu/edu/JetStream>

2.2 Monsoon Season and Weather Systems

Seasonal land-sea temperature differences caused by solar radiation leads to monsoon weather systems (Huffman et al., 1997). Onset of the monsoon results in the seasonal reversal of the wind, significantly influencing regions of western North Pacific and East Asia. In this section we present an overview of the winter monsoon and the associated weather systems over Asia.

2.2.1 General Synoptic Pattern of North East Monsoon

It has been recognized that the weather over western Northern Pacific and East Asia is significantly dominated by the reversal of wind and abundant rainfall during the change of the season (Chang et al., 2005). The East Asian Winter Monsoon (EAWM) defined between the months of November and March (Ding, 1990) is considered as one of the major components of the winter monsoon system over Asia. The EAWM also referred to as NEM, is the season that gives rise to north easterly winds in the lower troposphere. The intensity of the EAWM is reported to affect wide regions of eastern China, North & South Korea, Japan, and as well as the surrounding regions leading to profound social and economic impacts (Chang et al. 2006). The maximum heat that develops during the summer of EAM, starts moving over South Asia towards northern Australia region during the winter of EAM. During the winter season, northeasterly winds at low level prevail over a larger region north of the equator, extending from the Western Pacific towards the Indian Ocean. Although both these EAM monsoon circulations are consequences of thermal circulation resulting from the differential heating, the two monsoon systems differ widely in their behavior. For instance, the summer EAM has its heat source region much nearer to the equator (Krishnamurti, 1971), where the effect of Earth's rotation is small, whereas the winter EAM circulation surrounds a larger region in the tropics interacting intensely with regions in the extra tropics. Furthermore, the winter EAM is characterized by northeasterly winds at low levels associated with a high-pressure system with anticyclonic circulation over the Siberian Mongolian region. This anticyclone circulation

is an intense cold-core high pressure system, that is semi stationary in nature with its central pressure higher than any other pressure system on Earth (Ding, 1994) and leads to the outbreak of cold air masses which may encompass large areas from the western Pacific to East Asia and progresses rapidly southward as CSEs that affect the tropics.

Figure 2.4 below as reproduced from the COMET programme shows the general synoptic pattern for the NEM. The PJ is most evident during the winter season when the westerlies are at their strongest and have shifted to their southernmost point. This PJ merges with the SJ at a location close to Japan to create a broad, deep band of very high wind speeds over eastern Asia. In winter, this SJ is situated south of the Himalayas and stretches north eastward across south central China. It crosses the China coast near Shanghai and merges with the PJ in the vicinity of Japan. This confluence enhances cyclo-genesis in the SCS which probably initiates when the upper level trough accompanied by cold air masses and the positive vorticity advection approach the low level baroclinic zone (Prezerakos et al., 2005). The northerly winds develop due to a stronger high-pressure region over East Asia (regions of Korea, Mongolia, China, and Japan) that is sustained and intensified by cold air mostly from the Arctic. Winds turning anti cyclonically around the eastern extent of the Asiatic High and the associated weather lead to CSEs. These surges produce a strong and steady north to northeast wind along the Asia coast and are prominent near East China Sea, Yellow sea, Philippines and the SCS.

The strength of EAWM have been proposed to be based on several indexes which in turn are based on pressure gradients, wind shears, and winds. These indexes show a relationship with El Nino-Southern Oscillation (ENSO) in most of the studies (e.g., Chang et al. 2004; Chan and Li 2004). Moreover, some studies also reported a relationship with Arctic Oscillation (AO) (e.g., Wu et al. 2006; Jhun and Lee 2004; Gong et al. 2001). For AO, this relationship was interpreted based on correlations as a result of formation of snow cover over northeastern Asia and Siberia. For instance, when there is formation of snow cover during the AO positive phase, this indicates that there would be delay in the onset of EAWM and the weakened EAWM result in weaker cold advection

from high latitudes over the SMH region. Similarly, the weakening of the EAWM was also reported by Li and Bates (2007) during warm phase of ‘Atlantic Multidecadal Oscillation’. In addition, a dynamical perspective of EAWM weakening was proposed by Wang et al. (2009) which was reported to be related to an interdecadal variation of the quasi-stationary planetary waves.

Figure 2.4: General synoptic pattern for NEM.

Source: <http://www.meted.ucar.edu/>

The NEM is characterized by a dry season that dominates the northern Asian regions (e.g. northern Thailand, Vietnam) and a rainy season that dominates the southern Asian regions (Singapore, Malaysia, southern Thailand). CS or northerly surges are primarily associated with acute temperature drop, high winds and strong CS are also associated with freezing rain, severe frost, heavy snowfalls and sandstorms occur. During the NEM, the flow of the northerly wind is prominent along the coast of East Asia and is most prominent over the

SCS where it strengthens into a CS. The latter is considered as one of the most distinct extreme weather events that occur once or twice in a month and may last from a few days to one week or even longer (Zhang et al. 1997, Wu and Chan 1997, and Wu and Chan 1995). During their occurrence, they can also cause snowfall and heavy freezing precipitation (Park et al. 2011a; Chen 2002). The NEM reaches its maximum strength in January, and then starts to weaken in February and March as the Australian Low and Asiatic High weaken. It disappears completely in April coinciding with the northward movement of the Inter-Tropical Convergence Zone (ITCZ) as well as the reestablishment of the Tibetan High and Tropical Easterly Jet. During the NEM, the SCS region is under the influence of strong SH pressure system and active CS. In turn NEM has significant impact on the weather system over the coastal regions of SCS as described next.

2.2.2 Weather System of East Asia during NEM

“Among the monsoon systems around the globe, the Asian winter monsoon has the largest meridional domain that extends from the equatorial Maritime Continent and surrounding oceans to Northern Hemisphere mid-latitude Siberia (Ramage 1968; Chang et al. 2006)”. Therefore, in comparison to the Asian summer monsoon, the Asian winter monsoon has a stronger influence from the baroclinic systems in the higher and middle latitudes, the Maritime Continent and the equatorial heat source in the vicinity of northern Australia (Chang et al., 2011). Thus, the large scale atmospheric circulations associated with the Asian winter monsoon have considerably larger meridional span compared with the Asian summer monsoon. The extra tropical component of the EAWM i.e. the NEM is characterised by drier weather with low level winds blowing predominantly from the northeast associated with the anticyclonic system over SMH. The movement of the semi stationary anticyclonic system in south eastward direction i.e. approaching and advancing to China’s coastline and the western Pacific results in abundance of cold air masses that often occur and severely impacts the regions of China, Korea and Japan. The east-southward expansion of the surface SMH triggers cold air outbreaks and the stronger cold

air out breaks progresses very fast and within a very short time period it can reach deep into the tropics (Chang et al., 1983). Onset of the surge events originating due to the intensification of north easterly winds over the SCS is termed CS, Pressure surges or monsoon (wind) surges can lead to severe weather and heavy convection in the East Asia and the Maritime Continent region. Each winter, approximately 10 CS occurs in East Asia (Chen et al. 2004). These surges associated with intensified SMH exerts tremendous economic and societal impacts on the countries in East Asia. Their onset often causes snowfall and freezing precipitation over east Asian regions (Jeong et al. 2008; Chen 2002; Ding 1994; Boyle and Chen 1987) and are the root causes of the severe weather over the SCS (Chan and Li 2004). Since the strongest atmospheric circulations are located over East Asia, EAWM is considered as the highly influential planetary scale circulation system for winter over Asia. Embedded in EAWM are the severe weather disturbances that results in weather related disasters such as floods, torrential rainfall, cold spells and snow storms and have significantly contributed for the leading cause of damages over East Asia.

Two major temperature modes, namely the ‘northern mode’ and the ‘southern mode’ was reported by Wang et al. (2009) during EAWM. The northern mode was characterized by intensification of the Siberian High over the domain $55^{\circ}\text{N} - 70^{\circ}\text{N}$, $90^{\circ}\text{E} - 120^{\circ}\text{E}$, and a westward shift of the East Asian major trough. This mode represents a cold winter in the northern East Asia due to cold air intrusion from northeastern Siberia. The southern mode features strengthening of SMH over the domain $40^{\circ}\text{N} - 55^{\circ}\text{N}$, $90^{\circ}\text{E} - 120^{\circ}\text{E}$ and the deepened East Asian trough. This mode represents a cold winter and enhanced monsoon circulation due to cold air intrusion from Mongolia. Both these temperature modes show temperature spatial variability limited to not only East Asia, but rather covering the entire Asia. The northern mode does not show correlation with ENSO and is preceded by excessive autumn snow covers over southern Siberia. The southern mode shows correlation with La Nina episodes and is preceded by reduced autumn snow cover over northeast Siberia. These two temperature modes exhibit different spatial & temporal

structures and origin with respect to their inter-annual variations and with their structures being somewhat similar with respect to inter-decadal time scale.

The weather of SEA is driven by the Asian Monsoon System (AMS) which is considered to be one of the main circulation systems of the Earth's climate system. The SEA region primarily comprises two distinct geographic regions, a mainland section and a maritime section. It is characterised by complex terrain, land and water. As shown in Figure 2.5, the mainland regions consist of Peninsular Malaysia, Vietnam, Thailand, Laos, Cambodia and Myanmar. The maritime region consists of Singapore, Philippines, Papua New Guinea, Indonesia, East Timor, East Malaysia and Brunei. The SCS is considered to be a major component of the AMS because it is highly influenced by three monsoon systems viz. "Asian Australian Monsoon" system, "East Asian Monsoon (EAM)" system and "Western North Pacific Monsoon" system (Wang et al., 2009). Thus, for the research community, understanding the climate variability of SCS has always been a major challenge.

Figure 2.5: Map of SEA (reproduced from Ratna, et al. (2017)) with Country boundaries.

2.3 Modelling Extreme Weather Events

The frequent occurrences of the extreme weather events such as wind storms, tornadoes, hurricanes, floods, heat wave, cold wave and drought have severe impacts on societies and ecosystems as these events cause losses of life along with tremendous economic losses. An extreme weather event can be defined as a rare event occurring for a specific region and particular duration of the year (IPCC, 2007). Extreme weather events not only affect human health directly through e.g. prolonged cold and heat wave conditions but also indirectly by affecting e.g. food crops through floods, drought and pollution episodes. Furthermore, the extremes associated with precipitation and temperature also affect human comfort, tourism and energy consumption (Meehl et al., 2000). These extreme

events have been a challenge scientifically firstly because it is difficult to understand the complex mechanism behind their occurrence, and secondly because it is even more challenging to predict such events in the future. As per the IPCC definition, such events are rare in nature and follow different statistical laws than averages. Analyses of extreme events usually requires estimation of the probability of events that are more extreme in nature, i.e. those that have already been observed.

Many models in practice are being used to understand and mitigate extreme event scenarios. Practically three categories of models are used in practice. They include statistical/actuarial models (historical data is used to estimate the consequence of future events), physical models (the consequence of an extreme event is estimated from a scaled down model) and simulation or catastrophe models (includes pre-determined parameters and physical constraints for simulating events). Furthermore, the structure of these models may be classified into three classes, viz. diagnostic, investigative and predictive. Parameters to predict the extreme event can be derived from the set of observed data sets provided that the extreme events have occurred only within that data regime. However, if not then modelling the extreme event and hence its prediction would be difficult and challenging. Thus, statistically prediction of extreme events become difficult if data are incomplete. For instance, to predict the one in 200 years of extreme event (e.g. cyclone) available cyclone data is needed for a period of decades (e.g. 30 years). Even for simulation model the key parameters governing the extreme event may vary stochastically. Therefore, for ideal case in modelling and predicting extreme event a combination of statistical and physical approaches should be made. Further details on climate extreme analysis and validations are found in Tebaldi et al. (2006) and Fowler et al. (2007).

Climate extremes can broadly be placed under two general categories. The first category is based on climate statistics that occur every year such as yearly maximum rainfall or temperature. The second is based on an extreme event which does not occur necessarily every year at a given location such as floods and hurricanes. The approaches to study the

above extreme events can also be placed into three major categories; (i) statistical approach, (ii) empirical physical approach and (iii) numerical modelling (Garrett and Muller 2008). The statistical approach which is also referred to as extreme value theory, is purely based on fitting probability distributions (e.g., Gumbel 1958; Wilks 2006; Garrett and Muller 2008). The empirical physical approach is based on empirical observation and uses physical reasoning to provide a basis for an extreme value model (e.g. Wilks 2006). The numerical modeling approach uses a numerical model e.g. General Circulation Model (GCM) and executes the model for a very long period (e.g. Kharin et al. 2007) to allow for a detailed examination of the physics behind the extreme event for estimating the statistics of extreme events. However, as an extreme event is by definition, unusual and rare, the statistical quantification of potential change in their intensity and trends becomes a difficult task (Palmer & Raisanen, 2002). Thus, to elaborate on the extreme weather events as CSEs, numerical modeling is next reviewed.

2.3.1 Theory of Numerical Models

Numerical modelling based on full physics models is one of the most advanced and powerful methods to study the complex processes within the Earth System. These numerical models are thus mathematical representations of the Earth climate system along with its components including the atmosphere, hydrosphere, land surface, cryosphere, biosphere and others, depending on the level of complexity to be tackled. The most important motivation using the development and application of numerical model is the assessment of future climate changes. Recent research using climate models has become highly interdisciplinary and includes domains of physics (e.g. atmospheric sciences, ocean dynamics), chemistry (e.g. aerosols, kinetics) and biology (e.g. ecology) (Stocker, 2011).

The equations governing atmospheric flow are expressed in terms of nonlinear partial differential equations and are applied for simulating the behavior of the atmosphere. The meteorological elements include the primary variables such as temperature T , pressure p , density ρ , specific humidity q and the fluid velocity V (u eastward velocity, v northward

velocity, and w upward velocity). The Navier-Stokes equations that govern the motion of fluids are expressed as:

$$\frac{\partial \mathbf{V}}{\partial t} + \mathbf{V} \cdot \nabla \mathbf{V} + \frac{1}{\rho} \nabla p + 2\boldsymbol{\Omega} \times \mathbf{V} = \mathbf{g} + \mathbf{F} \quad (1)$$

Where $\mathbf{V}$ represents the fluid velocity, $\boldsymbol{\Omega}$ the Earth's angular velocity, $\mathbf{g}$ the gravity and $\mathbf{F}$ the friction term. The terms on the left hand side of the equation represents local acceleration, nonlinear advection, pressure gradient and Coriolis term. The terms on the right hand side of the equation represents gravity and friction.

The variables pressure (p), density (ρ) and temperature (T) are related through the Equation of State

$$p = R \rho T \quad (2)$$

where R is the universal gas constant.

From the energy conservation and the first law of thermodynamics, we have

$$C_v (dT/dt) + RT \nabla \cdot \mathbf{V} = Q \quad (3)$$

where C_v represents the specific heat at constant volume and Q represents the diabatic heating rate.

The equations of motion as represented through Navier Stokes describe the “conservation of momentum” while continuity equation represents the “Conservation of mass” in the atmosphere and is expressed as:

$$(d\rho/dt) + \rho \nabla \cdot \mathbf{V} = 0 \quad (4)$$

The conservation of water vapor substance is expressed as:

$$(dq/dt) = S \quad (5)$$

where S represents all sinks and sources of water vapor. The above system of equations shall provide a complete description of the evolution of the atmospheric variables once

appropriate initial & boundary conditions, external forcing, sinks and source information's are provided. The vertical component of the velocity becomes much smaller in comparison to its horizontal component for large scale motions. For such cases the vertical equation is replaced by a hydrostatic balance between gravity and vertical pressure gradient and is expressed as:

$$(\partial p / \partial z) + g \rho = 0 \quad (6)$$

Nonhydrostatic models are widely used nowadays while the hydrostatic models were used for the first fifty years of Numerical Weather Prediction (NWP). Initial development of Non-hydrostatic model focused on research for small scale meteorological phenomena nonlinear mountain waves or convection. Today, it has a much wider application and the models are developed and applied to numerical simulations and operational NWP by meteorological offices worldwide, e.g. UK Met Office, Japan Meteorological Agency, Deutscher Wetterdienst of Germany and others.

Numerical models typically divide the region of interest into a set of grid points where the complex mathematical calculations are performed. The density of such discrete grid points is highly dependent on the computing resources available. Thus, the area coverage by such models depends on the spatial and vertical resolution used. Numerical models which cover the entire globe, typically have a coarser resolution and are known as General Circulation Models. Regional models on the other hand are the numerical models which covers only the limited area, and which depends on initial and boundary conditions provided by the larger GCMs.

2.3.2 Global and Regional Models

There are two main types of GCMs that separately account for the changes within the atmosphere and the ocean. These are known as Atmosphere and Ocean General Circulation Models respectively. When the atmospheric and the oceanic components are coupled together to form a more complete system they are known as Climate models.

GCMs have originated from the Atmospheric General Circulation Models. GCMs subdivide the surface of the earth into a three-dimensional grid over the ocean and the atmosphere and consist of thousands to millions of grid cells as shown in the Figure 2.6. GCMs are being used for a wide range of applications to understand the Earth's climate system. They are used to reproduce observed features of past climate and to investigate potential future climate changes on a global scale, (IPCC, 2007). The most widely known and well recognized application of GCM lies in projecting the future climate state under various scenarios as related to the emission of greenhouse gases, e.g. carbon dioxide, that are provided by the Representative Concentrations Pathways. Whereas, Regional Climate Models (RCMs) are used for simulating regional climate at a higher resolution to better understand phenomena which are driven by small scale processes at the local level (Castro et al. 2005). As used RCMs typically have an atmospheric limited area model (Giorgi et al. 2001). RCMs require initial & boundary conditions which are typically provided by the GCMs that drive these RCMs as illustrated in Figure 2.6. Applications of RCMs include regional climate research, long term simulations and weather forecasting. The RCMs can further include other components as marginal lakes and seas which influence the regional climate. However due to computational limitations, regional climate modelling is generally constrained up to a spatial resolution of 25 to 50 km. They are successfully used today to provide a reliable projection of a regional climate (e.g., Lo et al. 2008) and provide reasonable predictors for the purpose of statistical downscaling at a high resolution (Guyennon et al. 2013). Reviews and detailed information on the performance of RCMs are available from papers by e.g. Foley (2010) and Rummukainen (2010).

Figure 2.6: Global & Regional Model domains (reproduced from *Intermountain West Climate Summary*, (2011))

As described above, the main purpose of RCMs is to add realistic small scale features to their driving data. In addition to the positive impact on small scale features, the RCM could also have a positive impact on the large scale circulation, improving this as well. (Laprise et al., 2008). However, since GCMs do contain errors in their large scales, improving the spatial resolution in the boundary conditions may be a prerequisite for improved regional features (Simmons, 2006, Diaconescu et al., 2007). Veljovic et al. (2010) also supported the idea of also improving the large scales and suggested that in addition to a better resolution in RCMs, an improved physical description in the RCM compared to the GCM can contribute to improvement of the large scales of the driving data via the use of two way nesting between the GCM and RCM.

2.3.3 Dynamical & Statistical Downscaling

GCMs are an important predictive tool to understand the future climate. However, they do not account for climate variability and change at a finer scale due to its coarser resolution which is typically 100-300 km. These representations while providing reasonable information for large scale climate variables (on average) yield unrealistic local climate information when applied directly to a regional scale. The coarser resolution of GCMs thus has major limitations in analyzing regional climate as coarser resolution of GCMs results in a smoothing of the land surface representation. This is because landscape features, viz. water bodies, mountains, land cover characteristics, infrastructure and components of climate system, coastal breezes and convective clouds occurs at a much finer scale less than 100 km. For decision makers addressing climate impacts, these heterogeneities are important as they drive potential impacts on e.g. crop production, and hydrology which are at a scale less than 50 km. Thus, GCMs require major further processing of their model outputs so that they can bring the global scaled model output to a more reasonable localized level. Such a process of obtaining localized information at a regional level is called “downscaling”. The downscaling processes provide a better information for regional climatic influences (such as local topography) at a finer scale by capturing the sub-grid scale contrasts and inhomogeneities. The downscaling processes can be further categorized into two separate approaches as dynamical downscaling and statistical downscaling. Of the two, dynamical downscaling extrapolates the climate information available from a GCM to a more localized region using high resolution grids of a RCM simulation. In addition to having large scale atmospheric information provided by GCMs via initial and lateral boundary conditions, these RCMs incorporate more complex topography, detailed description of physical process, land surface processes, surface heterogeneities to provide a more realistic climate at a spatial resolution of 50 km or less. However, as the RCMs are driven from GCMs, the overall performance of dynamically downscaled output from RCMs is subjected to the accuracy of the forcing of GCMs and the biases associated with it (Seaby et al., 2013). RCMs are further subjected to systematic errors and often require bias corrections.

Statistical downscaling technique in comparison uses various methods to determine the relationships between the climate pattern simulated from GCMs with that from an observed (e.g. measured) climate data sets. The process involves establishing empirical relationships using historical datasets for the climate variables of interest. Once determined and validated, the relationship is used to predict future climate variables of interest. When used with GCM future outputs, this approach has a critical assumption in that the relationship developed using historical climate is also valid under future climate (Zorita and von Storch, 1999). It remains to be proven if this holds.

Both dynamical and statistical techniques have successfully been used as reported in the literature. In addition, dynamical downscaling which employs regional climate modelling should provide better predictors or when coupled with statistical downscaling to provide higher-resolution output (Guyennon et al., 2013). Statistical downscaling requires thorough understanding of the climate variables at global and regional scale and thus more complex in establishing a relationship for both present and future climate but have the advantage of lesser computation requirements in comparison to the dynamical downscaling.

As described above, the generation of climate change estimates based on a downscaling using a cascade of modelling steps. All steps add uncertainty to the estimates of future climate change. For instance, Rowell (2006) and Déqué et al. (2007) studied the relative importance of the different sources of uncertainty for seasonal means of temperature and precipitation. They found that the major source of uncertainty is connected with the choice of GCM. For temperature and precipitation, the uncertainty of the RCM formulation and setup is comparable to the uncertainty associated with the emission scenarios. Räisänen et. al (2001) described four different sources of uncertainty; first, the uncertainty due to emission rate scenarios; second, the uncertainty due to GCM formulations; third the uncertainty due to the RCM formulations and set up; and fourth, the uncertainty due to internal variability in the climate system. They compared the results using 15 GCMs and concluded that model differences across GCMs create larger

uncertainty than that from internal variability of each model. A general discussion on the different aspects of uncertainty in climate modelling can be found in Foley (2010).

2.4 Cold Surges in South China

As discussed in Section 2.2, a cold air outbreak is considered to be one of the most severe weather events during EAWM, and which can be felt over a large spatial extent, stretching from East Asia to the western Pacific. These cold air masses coming from Siberia, Mongolia or regions of northern China progress southwards over China reaching its coastal areas and finally making further advancement into the SCS. When these outbreaks progress rapidly southward and impacts tropical regions, particularly in the vicinity of the SCS, it is referred to as a CS (Section 2.2). These occurrences of CS are associated with an acute temperature drop and increase in wind speed from the northerly direction. Sometimes, it is also associated with strong gusty winds, widespread rain and severe thunderstorm. The CS are frequent between the month of November and February and the more intensified surges further lead to strong convective activities over the SCS (Lu et al. 2007). The CS over the SCS is mostly confined to a shallow layer extending from the surface to 850 hPa pressure level (Ramage, 1971). While various definition of surge exists, the usual definition involves a drop-in temperature to several degrees Centigrade within a short period of time, and progression of a sustained period of strong northeasterly winds (e.g. Chan and Li, 2004). In South China and HK these events are also identified as “northerly surges”. Another type of surge is called easterly surge as described in Wu and Chan (1995), where the wind accelerates primarily in the zonal direction resulting in a sustained surface easterly winds. In contrast, the northerly surges are associated with a low level anticyclone moving eastward towards China’s coastline, and primarily affecting the northern SCS.

Extensive previous research on CS has been carried out using conventional meteorological data (e.g., Wu and Chan 1995; Chen et al. 2004; Zhang et al. 1997) as available from national meteorological agencies. Some recent literature have also focused

on studying the event using satellite data (Alpers et al. 2012). A favorable condition for the outbreak of CS is when the anticyclonic system over Siberia intensifies and the tropospheric disturbance over Lake Baikal in the form of wave-train (trough-ridge pattern) pattern is noticed. With the propagation of the wave-train south eastward and with the deepening of the trough over East Asia, northeast flow is intensified bringing cold air from Siberia towards Eurasia (Zhang et al. 1997). The amplified surface cold anomalies result in enhanced anticyclonic circulation leading to acceleration and expansion of the northwesterly flow (Chen 2002; Lau and Lau 1984; Joung and Hitchman 1982). Anticyclonic system over Siberia intensifies with the interaction between the surface cold anomalies and upper level wave train over Siberia towards the outbreak of CS (Takaya and Nakamura 2005a). However, some studies suggested that outbreak of CS are distinct from the wave-train pattern as discussed. A study by Takaya and Nakamura (2005b) demonstrated two different origins for the intraseasonal amplification of the Siberian High. First, a blocking from the Pacific and a second, a wave-train from the Atlantic. In addition to the above, previous literatures have also highlighted the influence of large-scale phenomenon such as Madden–Julian oscillation (MJO) (Madden and Julian, 1972) (Jeong et al. 2008), El Nino– Southern Oscillation (ENSO) (Chen et al., 2004), and Arctic Oscillation (AO) (Thompson and Wallace, 1998; Hong et al. 2008; Park et al. 2010, 2011a) on the outbreak of CSs. For instance, when AO is found to be negative, the deepening of the east coastal trough and intensification of Siberian High provides favorable conditions for the occurrence of CS (Jeong and Ho 2005; Gong and Ho 2004). Further details on cold air outbreak are available in Chan and Li (2004). Zhang and Xiao (2013) found that the intensity of the JS is intimately related to the atmospheric general circulation in mid-latitude East Asian region and significantly correlated with the EAWM.

The genesis of cold air outbreak starts with the intensification of the surface SMH pressure region and its subsequent southeastward extension. The center of this high pressure area may move either south eastward (Zhang et al., 1997; Ding and Krishnamurthy, 1987) or it may remain stationary while the cold air may propagate eastward as associated with smaller high pressure centers (Chan and Li, 2004). Previous

literature also indicates changes in circulations in the upper troposphere over Siberia could serve as precursors of cold air out break (e.g. Ramage 1971, Compo et al., 1999, Boyle and Chen, 1987). These works typically focus on a single model that describes a single CSE which evolves over a period of around one week. However, there are often consecutive signals of CS over the northern SCS where the surges appear to be series of sequential events with in-between surge duration of a few days (e.g., Dec 2005, Jan-Feb 2008, Jan 2011). Furthermore, all the individual surges are not always associated with the classical model discussed above in this Section i.e. splitting of the SMH system or having a substantial southeastward extension. The recent literature rather indicates that these surges can trigger strong convective activity with massive snowfall and intense rainfall as the cold continental air mass interacts with the warmer and humid maritime air mass from the SCS (Yang et al., 2010; Lu et al., 2010; Chen and Ding, 2007). Moreover, at local scale, the extreme weather events are triggered by a combination of nonlinear interactions between several environmental factors (Kolstad, 2015), including large-scale atmospheric conditions, location, topography and alignment (Webster et al., 2008; Subyani, 2011; Orr et al., 2014). Therefore, to analyze extreme weather phenomena at a horizontal length scales of few tens of kilometers requires a very high-resolution mesoscale model for resolving their complex dynamic and physical features (Powers, 2007; Webster et al., 2008). However, no detailed investigations using high resolution physical-based regional climate models have been reported or its use towards the understanding of mesoscale processes affecting the CS as what is presented in this PhD work. Nor is there reported work, to the author's knowledge on the role of coupled interaction between the three large-scale atmospheric circulations; the Siberian Mongolian High, the Aleutian Low and the Jet Stream on the evolution of CSEs as explained in this thesis.

Chapter 3

Model and Data Sets

3.1 Weather Research and Forecasting Model

One of the major challenges in simulating CSES in South China lies in developing a simplified numerical model with an appropriate atmospheric forcing and model configuration. A variety of choices has been adopted in previous studies and the issue is reflected by various authors on predicting extreme weather over South China using models, conventional data or satellite data (Chen et al. 2004, Alpers et al. 2012, 2015). It is our explicit aim here is to create a simplified representation of model configuration and atmospheric forcing that represent the key aspects of the complicated physics associated with the CSEs in South China within the context of mesoscale and synoptic meteorology. In this chapter, a brief review of Weather Research and Forecasting Model (WRF) model along with a discussion of the model physics, nesting technique and forcing used by the author to simulate CS in South China is presented. The chapter will contribute towards the final formulation of the forcing to be used in chapter 4.

3.1.1 Introduction to Weather Research and Forecasting Model

The WRF model is a Numerical Weather Prediction (NWP) system designed specifically for research in atmospheric sciences with applications for real time forecast. This WRF system is developed by NCAR in collaborations with the leading organizations such as National Centers for Environmental Prediction (NCEP), University of Oklahoma, Air Force Weather Agency, National Oceanic and Atmospheric Administration, Forecast System Lab and Federal Aviation Administration. WRF's dynamics are represented in its atmospheric fluid flow solvers, or cores. In the early 2000s, another solver, the Nonhydrostatic Mesoscale Model (NMM) core from NCEP (Janjić et al. 2001; Janjić 2003) was added to provide an alternative solver option. Each dynamic core corresponds

to a set of dynamic solvers that operate on vertical coordinate system, grid staggering and a particular grid projection. The two WRF variants are called the Advanced Research version of WRF referred to as WRF-ARW and the other, the Non-Hydrostatic Mesoscale Model of WRF referred to as WRF-NMM. The dynamical core includes modelling of various effects such as advection, diffusion, buoyancy, Coriolis, pressure gradients and others. Both these cores are non-hydrostatic “Eulerian mass” dynamical cores that enable the mathematical equations to be used over a wide range of atmospheric scales.

WRF is a “Community Model” i.e. it is freely available in the public domain. As such the WRF is widely used in areas spanning atmospheric research, data assimilation, coupled model applications and real-time forecasting. They are designed to be flexible and efficient to work on parallel computing platforms for various applications to allow spatial resolutions ranging from kilometres to even metres. The major components of the WRF software infrastructure is shown in Figure 3.1a. The ARW core has an advantage over NMM as it can be used for performing additional applications on regional climate research, chemistry applications and idealized simulations at different scales. The key features of ARW include WRF ARW dynamic solver, WRF data assimilation (WRFDA) and numerous physics packages. The WRF-ARW modelling system components consisting of software infrastructure, key features of ARW, postprocessor utilities and visualization tools are shown in Figure 3.1b.

In running WRF, the WPS pre-processor is primarily used to define the simulation domains (single or nested); to interpolate the terrestrial data sets (e.g. terrain data, soil layers and land use categories) over the domain; and to interpolate the meteorological data from the GCM over the WRF domain. In the model system, the WRFDA programme is primarily used to ingest observations such as radiance, radar and precipitation when used for real time simulations. This is further used for updating the initial conditions when the model is set up for a run in a cyclic mode. The ARW solver is the most important component of the entire WRF-ARW modelling system. It integrates the compressible, Euler’s nonhydrostatic equations based on a terrain-following hydrostatic pressure system

(Laprise, 1992) when the surface has topography. It encompasses the initialization routines, numeric/dynamics options, physics schemes, and data assimilation packages.

Figure 4.1: WRF (a) Software Infrastructure (b) Modelling System Flow Chart.

Source: <http://www2.mmm.ucar.edu/wrf/users/model.html>

Using Ooyama (1990) philosophy of formulating the prognostic equations in terms of variables that have conservation properties for both the mass and height coordinates, the equations are represented in flux form. Figure 3.2 represents the ARW vertical coordinate. In the figure, P_h represents the hydrostatic component of the pressure, whereas P_{ht} and P_{hs} indicates the value along the top boundaries and surface, respectively. The value for η ranges from a lower value of 0 at the upper boundary to a higher value of 1 at the surface of the model domain. Terrain-following coordinates also called sigma coordinates, and as merged with pressure surfaces is called a hybrid sigma-pressure coordinate. In the WRF model, this coordinate system is defined in reference to the dry air mass, and so $\eta = (P_{hd} - P_{ht}) / \mu_d$ where P_{hd} is the hydrostatic pressure of the dry atmosphere and $\mu_d = (P_{hs} - P_{ht})$ represents total mass of dry air in the column.

Figure 3.2: The ARW model terrain following hydrostatic vertical coordinate, adopted from Skamarock et al. (2008).

The use of a terrain following coordinate system reduces the complexity in implementing the lower boundary condition over the terrain (Wu and Arakawa, 2011; Laprise, 1992).

There are, however, disadvantages of using the terrain-following coordinates in regions of steep terrain where it accounts for large error for computing horizontal pressure gradient (Janjic, 1989). These errors can lead to the generation of spurious precipitation. Newer methods have recently been developed to reduce the erroneous circulation caused by such pressure gradient biases, such as the smoothing of topography in terrain-following coordinates or methods of implementing topographic boundary conditions in height coordinates (Klemp, 2011; Wu and Arakawa, 2011).

The WRF model currently supports four different map projections; Lambert conformal, Mercator, Polar stereographic and Latitude-Longitude (global domain). In this study, the Mercator projection which is also known as cylindrical orthomorphic projection is used. The time integration is a time-split scheme, whereby high-frequency modes are integrated over a smaller timestep than lower frequency mode. This technique thus reduces the computational time that would be required whilst maintaining numerical stability. The ARW solver uses the Runge-Kutta time integration schemes of second and third order with the capability of adaptive time stepping. ARW uses the staggered Arakawa C grid structure wherein the flow variables (u , v) are defined along the cell faces of the domain and the thermodynamic variables are defined at the center of the domain. The flow variables are staggered one-half grid length away from the center of the domain where the thermodynamic variables are located. The WRF model is capable of executing either one-way nesting or two-way nesting. The use of nesting allows for simulations of multiple computational domains of higher resolution grids within the larger domain. The lateral boundary conditions needed for the finer-resolution grids “child” domain, are interpolated and provided from the coarser-resolution grid, “parent” domain. In a one-way nested simulation, there is the only exchange of information from the “parent” domain to the “child” domain. In two-way nesting both the finer and coarser grids are integrated simultaneously by additionally aggregating information from the child domain and feeding the information back onto the parent domain.

The WRF has been widely used to dynamically downscale from the synoptic and larger scale atmospheric circulation in order to provide a high-resolution analysis of weather and climate in regions of complex terrain. The model has been widely used for a variety of different applications and have been evaluated against observations in different regions, including, for instance, Europe (Tuccella et al., 2012), North America (Yver et al., 2013) and east Asia (Gao et al., 2014). Gao et al. (2015) performed dynamic downscaling assessment of summer temperature and precipitation in East Asia using WRF model. Their study provided an assessment of the model sensitivity and its ability to produce climate features in East Asia. Chotamonsak et al. (2011) showed that WRF successfully reproduced the observed spatial distribution of temperature and precipitation over SEA for the period 1990-1999. The future projection (2045-2054) from WRF model showed that projected warming varied from 0.1 to 3 °C depending on the season and with increased averaged precipitation. Other studies using the WRF model include investigation of the solar radiation parameterizations and ozone absorption during the summer of EAM (Kim et al. 2011), features of a severe dust storm and its transport characteristics in East Asia (Huang et al. 2013) and others. However, one of the main issues for WRF is the large amount of computing time needed to achieve fine scale resolution. A way to address this issue is to combine with statistical methods including the use of regression models and stochastic weather generators or weather typing, as documented by Wilby and Wigley (1997), Abatzoglou and Brown (2012).

For simulating regional climate, the choice of different parameterization schemes in RCM vary in their performance. For instance, Huang et al. (2011) reported superior performance of the WSM3 microphysics scheme over the others in WRF model for simulating torrential rainfall event related to the large-scale weather system in Xinjiang region during 25–26 May 2009. More recently Zhu et al. (2014) reported that WSM3 was the best scheme among eight microphysics options i.e. Lin, Kessler, WSM3, WSM5, and WSM6, Eta, WDM5, WDM6 for simulating the light to heavy rainfall and downpours of a heavy rainfall event that took place in southern China during 6–7 May 2010. For simulating three events of intense precipitation over the southern Peninsular Malaysia in

the winter monsoon of 2006–2007, Ardie et al. (2012) tested four types of cumulus parameterization schemes in the WRF model. In their simulation the BMJ scheme showed high errors during the first event but showed better performance for the second and third events. They found that GD (Grell-Devenyi) and KF (Kain-Fristch) schemes performed weaker in comparison to BMJ (Betts-Miller-Janjic) which simulated well higher values for rainfall. The effect of different cumulus schemes in simulating typhoon track and intensity over Northwest Pacific Ocean was studied by Li (2013). The result showed that only KF scheme could simulate the most severe typhoon while GD and BMJ schemes could simulate only weaker typhoons. Since, there are several different physical and dynamical schemes which are used in simulation there is always a controversy surrounding any perceived advantage of one particular scheme over others. Based on the above configurations on the choice of parameterization schemes, a series of numerical experiments were conducted to identify the more reasonable configuration for simulating the three CSs onset in South China. To investigate the sensitivity of cumulous parameterizations scheme during the onset of CSs, we conducted experiments using different parameterization schemes available in WRF. The results of simulations and discussions are presented in Appendix A. Our analysis showed that GF scheme comparatively performed better in capturing the onset of CSs (Figure A.1, Appendix A).

The WRF model version V3.7 released on 17th April 2015 (<http://www2.mmm.ucar.edu/wrf/users/model.html>) is used in this PhD work. The model was successfully installed, configured and ran at the NTU's High Performance Computing Centre (HPCC). Additional information on NTU's HPCC is available at <http://www.ntu.edu.sg/cits/hpc/Pages/default.aspx>

3.1.2 WRF Physics

The WRF model offers multiple physics options which can be tuned to meteorological conditions for the specific region of study, enabling users to optimize WRF for specific scales, geographical locations and applications. The physics options include microphysics, cumulus parameterisations, land surface schemes, surface schemes, planetary boundary layer schemes, and radiation schemes. The physics and interactions contained within each of these categories are briefly described below, with emphasis on those schemes used in this PhD work. Further details on all schemes are available from Skamarock et al. (2008). The microphysics modules model explicitly precipitation, cloud and grid resolved water vapour processes. The main difference between the various microphysics schemes is the number of hydrometeors which are rain, cloud water, cloud ice, water vapor, snow and graupel. All schemes include rain, cloud water and water vapour with more sophisticated schemes additionally including some or all solid water phases of cloud ice, snow, graupel and hail. The interaction of solid and liquid water, such as the accretion of super-cooled water onto snow particles to form graupel results in the Mixed phase processes. Figure 3.3 shows the difference in the number of hydrometeor interactions between a scheme with only warm phase interactions and those with mixed-phase processes. For simulations with a horizontal resolution of greater than 10 km for which updrafts can be resolved and also in areas where deep convection is common, it is recommended to use a scheme that includes mixed-phase processes (Skamarock et al., 2008).

Figure 3.3: Schematic depicting the different interactions between a microphysics scheme with (a) only warm cloud processes and (b) mixed cloud processes. Q_r is rain water, Q_c cloud water, Q_v water vapour, Q_g graupel, Q_i cloud ice and Q_s snow.

Source: <http://derecho.math.uwm.edu/classes/NWP/10-Microphysics.pdf>

The other significant difference amongst the microphysics schemes is between double-moment and single-moment schemes. Double-moment schemes explicitly calculate number concentrations in addition to the mixing ratios as calculated by traditional single-moment schemes. Use of a double moment scheme tends to decrease the excessive light rain often produced by model's microphysics (Lim and Hong, 2010). Two such double moment schemes within the WRF model options are the WRF Double-Moment 6-class scheme and the Morrison 2-moment scheme. Both include all mixed-phase processes and thus have six hydrometeors. The WRF Double-Moment 6-class scheme computes two moments (i.e. the mixing ratio and number of concentration) for cloud water vapour, rain, and cloud condensation nuclei, whilst the Morrison double-moment scheme computes two moments for cloud ice, cloud rain, graupel and snow. This suggests that the Double-Moment 6-class scheme may perform better for warm processes involving mostly cloud water and rain. Conversely the Morrison scheme may produce better results for colder processes that involve larger amounts of snow, ice and graupel. The vertical distribution of hydrometeors also plays a key role in directly affecting the precipitation rate through

the processes in the microphysics scheme. It is important to note that any error in the hydrometeor distribution may account for large biases while predicting rainfall. Thus, for improving the simulation associated with dynamics of the convective processes, it is of immense importance to analyse cloud and precipitation at various model resolutions. For instance, during deep convection, cloud parameterization is considered when the moist static instability exists over the entire atmospheric column while for shallow convections it is considered even when the moist static instability is present only over some parts of the atmospheric column.

In contrast to the microphysics scheme, cumulus parameterisations are used for simulating the sub-grid scale effects of shallow clouds and convective activity. The parameterisations represent the unresolved downdrafts and updrafts along with the compensating motion outside the clouds and subsequent vertical fluxes. They are designed for resolutions lower than 10 km where the assumption of convective eddies being entirely sub-gridscale is valid. They have also been shown to be helpful in triggering convection at higher resolutions (Lowrey and Yang, 2008). The available cumulous schemes vary in complexity and design. Within the WRF model, precipitation produced by processes explicitly resolved is classified as grid scale rainfall, while precipitative activity at length scales smaller is classified as cumulus rainfall and calculated within the convective parameterisation. Care must be taken in the implementation of this separation, as it is due to spatial resolution and does not represent a distinct physical difference in how the precipitation is formed. Convective cells are often associated with intense short-lived precipitation (Thomas et al., 2010) and thus the cumulus rainfall is often assumed to represent precipitative activity that is heavier and shorter-lived than the grid scale rainfall. The Grell–Freitas ensemble convective cumulus scheme [Grell et al., 2014, based on a stochastic approach originally implemented by Grell and Devenyi (2002)] and modified for use in higher resolution has found widespread use.

The model physics used in the present PhD work is shown in Figure 3.4. The atmospheric inputs to the cumulus scheme are surface velocities and heat and moisture exchange

coefficients calculated in the surface layer scheme, precipitation from the microphysics and convective schemes and radiative forcing from the radiation schemes. The land surface model (LSM) calculates surface moisture and heat fluxes ((latent heat (LH) and sensible heat (SH)) over land (Figure 3.4) using atmospheric variables along with the land's state variables (e.g. soil moisture and temperature) and land surface properties (land-use, vegetation and soil type). The moisture and heat fluxes then provide the lower boundary conditions for the vertical transport in the planetary boundary layer scheme. The LSM also produce model outputs of surface and subsurface runoff. Various schemes treat vegetation, root zones and canopy effects differently, and have varying levels of sophistication in modelling moisture fluxes and thermal in multiple soil layers. The Noah LSM (Chen and Dudhia, 2001a) has four levels of soil temperature and moisture along with snow cover and canopy moisture prediction. This LSM includes root zone depths as dependent on vegetation type, evapotranspiration, soil drainage and runoff and further include the effects of vegetation categories (based on the MODIS land-cover classification of the International Geosphere-Biosphere Programme), monthly vegetation fraction and soil type. The Rapid Update Cycle (RUC) LSM model has a default of six soil levels, which can be set to a higher number. This is a more sophisticated scheme to the Noah scheme for cold climates as it has a multi-level snow model. Computationally however, it is much more expensive as it requires a smaller timestep to maintain numerical stability. At present, no lateral interaction between grid points in any of the LSMs is available in the WRF model. A lateral router has been developed for the Noah model (Gochis and Chen, 2003), however this is computationally expensive. The Noah LSM also has biases in simulating snowmelt and runoff (Slater et al., 2007; Bowling et al., 2003). Thus, in this PhD work the newly developed Noah Multi-parameterization (Noah-MP) Land Surface Model with snow-cover prediction and canopy moisture (Niu et al., 2011) has been used as it better correlates with the observed values of soil temperature, sensible and latent heat fluxes, soil moisture and albedo (e.g. Gao et al. 2015, Chen et al. 2014).

Figure 3.4: WRF physics interactions. Model physics parameterization schemes valid for South China, HK.

The surface layer schemes are designed to represent the interaction between the land surface and the atmospheric surface layer between the first 50-100 m of the atmosphere. The surface scheme calculates the velocity shear caused by surface friction, which is dependent on the roughness of the land surface. Heat, moisture and momentum fluxes in the layer are also calculated. At present each surface scheme is related to a particular

planetary boundary layer (PBL) scheme, and thus these must be chosen together. The PBL schemes are responsible for determining the flux profiles of moisture, heat and horizontal momentum within the well-mixed PBL, as well as within the sub-grid scale for the entire atmospheric column. The boundary layer has been defined as the atmospheric layer that is directly influenced by the Earth's surface and which responds to surface forcing of friction, heating and cooling on timescales of less than a day. This is where significant fluxes of heat, mass or momentum are transported by small-scale turbulent motion (Garrat, 1992). The height of this layer varies both spatially and temporally as dependent on surface heating and cooling and the presence of clouds (Garrat, 1992). It can vary from around 50-100 m to a few kilometres. However, this height as defined within a model varies across models and the different planetary boundary schemes within WRF have different definitions for the boundary layer top. All the surface layer schemes are one-dimensional and are based on the assumption that a clear scale separation exists between the resolved and sub-grid scale eddies. This is valid for grid sizes above a few hundred metres. In this PhD work, the Yonsei University Scheme (YSU) PBL is used. This scheme from Hong et. Al (2006) has been modified in WRF version 3. The critical bulk Richardson number in this version has been increased from zero to 0.25 over land to allow enhanced mixing in the stable boundary layer (Hong and Kim 2008).

The effects of atmospheric radiation are calculated by separate schemes for the incoming solar shortwave radiation and for the longwave radiation emitted from gases and surfaces. The scheme calculates the vertical profiles of the atmospheric heating considering the effect of radiative absorption and provide surface downward fluxes for the ground heat budget. Reflection, absorption and scattering are included and the radiation responds primarily to simulated water vapour distributions and clouds as well as ozone concentration and carbon dioxide. This assumption is valid provided that the horizontal grid-spacing is far larger than that in the vertical. To maintain validity of this assumption, the vertical resolution must be increased when horizontal resolution is increased significantly. The Rapid Radiative Transfer Model models longwave processes due to

water vapour, ozone, carbon dioxide and trace gases, as well as accounting for cloud optical depth. It calculates radiative forcing in 16 spectral bands. There are options for shadowing effects and terrain slope on the surface solar flux and different options can be selected to produce the best results for the variables of interest in a particular region and climate.

The Advanced Research WRF (ARW) dynamical solver integrates non-hydrostatic fully-compressible Euler equations, corresponding to the Navier-Stokes equations with zero viscosity and heat conduction terms, along with equations for conservation of mass, energy and moisture. These equations are discussed in Section 2.3.1. As discussed in Section 3.1.1, the use of the hybrid vertical coordinate system reduces the erroneous circulation associated with the pressure gradient biases that leads to spurious precipitation. Within the WRF model the primitive equations have a different form from Section 2.3.1, due to their expression on this vertical coordinate system, the inclusion of moisture, transformation due to the map projection chosen, and the use of a perturbation form. For reducing truncation errors in computing horizontal pressure gradient and for reducing rounding errors in computing buoyancy and the vertical pressure gradient, the WRF model makes use of perturbation variables. This use of perturbations requires a reference state. In the WRF model the hydrostatic state is considered as a reference state and is purely a function of height only (Skamarock et al. 2008). WRF model also uses a split-time scheme, whereby high-frequency modes are integrated over a smaller timestep than lower frequency mode and has the capability of adaptive time stepping. The time step chosen for the low frequency integration is based on the Courant number given by $Cr = (u \cdot \Delta t) / \Delta x$ for each of the wind components (u,v,w). A Courant number of 1.0 means that all atmospheric particles in one grid box of size Δx , will travel to a neighbouring grid box in a timestep Δt . The timestep is altered for each integration in order to obtain a Courant number as close to, whilst still less than, the specified Cr_{target} (Target maximum Courant number) with, $1.1 \leq Cr_{target} \leq 1.2$ being used.

It is noted there are many permutations of model option choices and the computational time to test all of these is extremely large. Some regions and variables will be more sensitive to particular options. For example, Lowrey and Yang (2008) showed that the simulation of an extreme precipitation event in central Texas is most sensitive to the cumulus scheme, slightly sensitive to the microphysics and almost unaffected by the radiation schemes. In tuning a model to a particular region care can be taken to test only those options that the variables of interest are likely to be most sensitive to. In this PhD work, the model physics are based on a series of sensitivity experiments carried out for the SCS region and published in Kumar et al., (2016). The physics options used (Figure 3.4) include the Stony-Brook University microphysics scheme (Lin et al., 2011), Grell–Freitas ensemble convective cumulus scheme (Grell et al., 2014), RRTMG shortwave (SW) and longwave (LW) radiation schemes (Iacono et al., 2008) and Rapid Radiative Transfer Model (RRTM) longwave radiation, the Yonsei University PBL scheme (Hong et al., 2006), and the Noah Multi-parameterization (Noah-MP) Land Surface Model with snow-cover prediction and canopy moisture (Niu et al., 2011).

3.2 WRF Forcing Data & Data sets

As required in running limited area models, the WRF model requires the specification of initial conditions for the entire domain and lateral boundary conditions at regular intervals (3-12 hourly). These initial and lateral boundary conditions are user-defined and interpolated from lower-resolution forcing data. The three-dimensional fields required by the WRF model are temperature, relative humidity, geopotential height and horizontal wind fields. Time-dependent two-dimensional required fields are surface and mean sea-level pressure, skin temperature (temperature of surface layer of the Earth), surface temperature, surface relative humidity and surface winds. SST can be provided either as a static or time-dependent field. If the Noah LSM is used, then layers of soil temperature and moisture are also required. Snow depth and a sea ice may be included if relevant. The WRF model also requires some 2D static fields such as albedo, terrain elevation, vegetation type and soil texture. These data are all horizontally interpolated onto the

projected domain. For the vertical extrapolation near the surface needed if the WRF high resolution topography is lower in elevation than that of the forcing data, the horizontal winds and relative humidity are assumed to have zero vertical gradient and temperature has a default lapse rate of -6.5K/km .

For lateral boundaries, WRF model provides a choice between periodic, open, symmetric or specified conditions. Periodic, open and symmetric conditions are generally used for idealised simulations while real simulations require specified conditions using data from forecasts (e.g. GCMs) or gridded observations (e.g. Reanalysis). The boundary condition specification further comprises of both specified and relaxation zones. The latter is referred to as a nudging or relaxation boundary condition (Skamarock et al., 2008). For the specified zone, the meteorological variables are entirely specified by the forcing data at the edge of the domain. This zone is typically only one grid wide, but this width can be increased to strengthen the influence of the forcing data. For relaxation or nudging the weighting of the nudging then decreases to zero through the width of the relaxation zone. The default relaxation zone width is 4 grids wide, though there is an option to change this thereby increasing or decreasing the large-scale influence. All nested domains use specified lateral boundaries. The specified boundary conditions apply to the potential temperature, horizontal wind components, geopotential, pressure and water vapour. Excluding water vapour, which is specified, all microphysics variables and scalars have flow dependent boundary conditions within the specified zone. One other option for real simulations is the tropical belt configuration where the model domain is completely wrapped around the equator. In this case the lateral boundaries are a combination of specified in the y direction (north-south) and periodic in the x direction (east-west). This configuration is useful for the study of waves travelling in the east-west direction (e.g. Tulich et al., 2011).

3.2.1 Reanalysis

Reanalysis is the name given to the climate data output or product using observational data assimilation systems based on recorded atmospheric observations, (Bengtsson and Shukla 1988; Trenberth and Olson 1988). Such systems comprise a series of analysis steps in which forecast model information for a short time period, typically around 6 hours, is combined with observations for that period to provide an estimate of the atmospheric state. As observations of all standard atmospheric fields at 6-hourly resolution for all points over the globe do not exist, the model forecast state provides a first approximation for such missing fields. This approximation may further be improved by extrapolation of near-by observations. Included are land surface observations, data from various atmospheric profilers such as radiosondes, flight-level data from aircraft, surface data from buoys and various forms of satellite data. The forecast model is a regional or global model in which the evolution of the atmosphere is governed by physical equations as described in Section 2.3.1. The incorporation of observational data is intended to force the model towards the actual meteorological state of the atmosphere for each time period. Reanalysis data are designed to be the closest available estimate for a complete gridded observational dataset. Different reanalyses have been developed by various institutions. Improvement in both the background forecast model and the assimilation techniques has occurred over time. These data can be broadly categorised into three major phases; the period from 1950 to 1957, which is the evolution of the observing system; the period from 1958 to 1978 with the modern radiosonde network; and the era from 1979 to present with the modern satellite era. The ERA (ECMWF Reanalysis) -15 was the first reanalysis projects produced by European Centre for Medium-Range Weather Forecast (ECMWF) around 1995 (Uppala et al., 2005). Later products from the ECMWF include the ERA-40 and ERA Interim reanalyses, whilst the NCEP/NCAR produces two reanalysis projects. The first is the NCEP/NCAR R1 (Reanalysis 1) where the data spans from 1948 to present and the second product R2 is an update where the data spans from 1979 to present. Further details on the products can be found in Kanamitsu et al., (2002). Due to differences in the forecast models and

assimilation techniques, reanalysis products can have significant variations in some fields (Marques et al., 2010; Grotjahn, 2008). Additional datasets and the calibration of new satellites can further introduce variations in the reanalysis. It is noted that while data assimilation methods are not designed to correct bias during the analysis step nor does it conserve global atmospheric mass, energy or angular momentum, the use of a model does at least enforce physical coherence between variables. Reanalysis can, if used with care, be an extremely useful tool in atmospheric physics, particularly for use as forcing data for limited area models such as the WRF model.

Though reanalysis data sets are the most widely used data sets in modern atmospheric science, they do suffer from inhomogeneities as well as distributional and serial dissimilarities (Kistler et al. 2001; Trenberth et al. 2001; Bengtsson et al. 2004; Sterl 2004; Trenberth and Smith 2005; Pitman and Perkins 2009; Screen and Simmonds 2011; Trenberth et al. 2011). These are especially frequent in regions where the observational network is sparse and have been interchangeably used and implicitly assumed to be “perfect” in downscaling studies (Jones et al. 2011). This is an issue of some concern, as the results of both dynamical and statistical downscaling approaches have been shown to be sensitive to the choice of the reanalysis data used for their development (Fernández et al. 2007; Koukidis and Berg 2009).

3.2.1.1 NCEP FNL Data sets

The Final (FNL) data sets are the product from the Global Data Assimilation System, produced by NCEP’s model, the Global Forecast System (GFS). The model was run at 12-hour intervals to produce multiple-day weather forecast from April 1997 through June 2007. FNLs are analyses that are produced after the model has completed the forecast cycle, and they include the most complete set of observations available for that given cycle. The FNL from a given cycle, which represents the best available analysis, is then used in the initialization of the next cycle. FNL data are available on pairs of 2.5-degree hemispheric grids (North and South) at the Earth's surface and sixteen vertical levels from

1000 millibars up to 10 millibars. Parameters include sea level pressure, surface pressure, SST, temperature, geopotential height, relative humidity, potential temperature, precipitable water, meridional & zonal winds, and vertical wind speed. In this PhD work, the WRF model is driven via initial and lateral boundary conditions using the NCEP FNL datasets available at $1.0^0 \times 1.0^0$ grid resolution and 6 hr time intervals.

3.2.1.2 ERA 40 Data sets

The ERA-40 reanalysis dataset (Uppala et al., 2005) is provided by the ECMWF. The data sets span over a period of 45 years starting from 1st September 1957 to 31st August 2002. Wang and Yang (2008) showed that different reanalyses can result in very different outputs from the WRF model as primarily caused by differences in the flux of water vapour across the lateral boundaries. The ERA 40 data were used for the analysis of CSEs before 1979 where ERA Interim data are not available. Known problems with the ERA-40 data include excessive precipitation over tropical oceans and unrealistically high values of global average precipitation compared to evaporation (Uppala et al., 2005). The ERA-40 reanalysis was designed to improve upon the previous ERA-15 reanalysis and include numerous additional observational datasets. The model has 60 vertical levels and uses a spectral resolution of T159, corresponding to a spatial resolution of approximately 125 km with variables available at 6-hourly intervals. However, there were several inaccuracies exhibited by ERA-40 that included too-strong Brewer-Dobson circulation in the stratosphere and too-strong precipitation over oceans. These inaccuracies are eliminated or significantly reduced in ERA Interim. Since ERA Interim extends from 1979 to the present, as such this PhD research uses ERA 40 as the analysis of CSEs before 1979 and ERA Interim after 1979.

3.2.1.3 ERA Interim Data sets

Also used in the PhD work is the ERA Interim reanalysis dataset (Dee et al., 2011) as provided by the ECMWF. This dataset covers the period from 1st of January 1979 till

present and is continually updated in near real time. The ERA Interim reanalysis provides improvements on the previous ERA-40 dataset, including improvements to the forecast model which significantly reduces the stratospheric biases between model and observations. The ERA Interim data assimilation system uses of the 4-dimensional variational analysis, Integrated Forecast System (IFS release Cy31r2) spanning the twelve-hour analysis window. The model comprises of 60 vertical levels and has a spectral resolution of T255, corresponding to a spatial resolution of approximately 79 km and variables are available at 6-hourly intervals. Further details on the product is available in Dee et al. (2011). In this PhD research ERA 15 is used for the detailed analysis of CSEs after 1979.

3.2.2 Satellite Data sets

The recent Advanced Scatterometer (ASCAT) dataset is used in this PhD work for comparing the WRF predicted wind fields over SCS and near the coast of HK. The datasets comprise of the measurements for surface wind vectors available at 10 m height above the ocean surface. The product is made available either with a 50 km resolution having a 25 km grid spacing or with a 25 km resolution having 12.5 km grid spacing. This remotely sensed surface wind field have been widely validated for its accuracy of ocean surface wind vectors (e.g. Bentamy et al. 2008; 2012; Sluzenikina et al. 2011). The main application of the wind products lies in operational nowcasting due to its high spatial resolution (Von Ahn et al. 2006). Detail information on ASCAT product is available in the product manual at the web link

http://projects.knmi.nl/scatterometer/publications/pdf/ASCAT_Product_Manual.pdf

3.2.3 Observational Data sets

Observational weather station data comprising of daily mean values of the meteorological variables, including mean sea level pressure, wind speed, minimum temperature, and total daily rainfall were available for the HKO station (22°18'07''N; 114°10'27''E) and the

Waglan Island station (22°10'56''N; 114°18'12''E) as downloaded from Hong Kong Observatory (HKO). Detailed information on HKO data sets are available at <http://www.hko.gov.hk/publica/pubdata.htm>. In addition, observational weather charts archived by HKO at 6 hourly intervals have been used in the study.

Chapter 4

Surge Analysis and WRF Modeling for South China and Hong Kong

4.1 Introduction to the CSEs of 2008, 2009, and 2016

The arrival of a CSE is generally characterized by a strengthening of northerly surface winds, a sharp drop in surface temperature, and a steep rise of surface pressure. Among the synoptic disturbances, CSEs as characterized by a succession of cold air outbursts from the Siberian High (described in Section 2.4) is considered as one of the most impactful weather events for the East Asia region with the sharp temperature drops potentially triggering disasters as reported by (Chan and Li 2004; Chang et al. 2006). As reported by Chen et al. (2002), the onset of CS in East Asia when the anticyclonic system over Siberia intensifies and the tropospheric disturbance over Lake Baikal in the form of wave-train (trough-ridge pattern) pattern is noticed. With the propagation of the wave-train southeastward and with the deepening of the trough over East Asia, northeast flow is indicated bringing cold air from Siberia towards Eurasia (Zhang et al. 1997).

Various criteria have been proposed in the literature to indicate the onset of a CS though none has been universally accepted. CS onset in HK is defined by the HKO as (i) temperature drop $> 2^{\circ}\text{C}$ during the past 24 hours or (ii) mean temperature during the next 6 hours being less than the 24-hourly mean temperature recorded 30 hours ago by $> 2^{\circ}\text{C}$. A 2°C drop in daily mean temperature within a 48-hour period has been earlier proposed to be a criterion for the arrival of a monsoonal CS (Chu, 1978; Lam, 1979). Riehl (1968) earlier proposed a large pressure difference between 30°N , 115°E (i.e. just north of HK) and HK as exceeding 10 mb to be the surge forecast criterion for the arrival of a northerly CS. More recently, Chang et. al. (2005) devised a CS index in SCS using the average 925

hPa meridional wind between 110°E and 117.5°E along 15°N. A CS will occur when this index exceeds 8 m/s.

Three recent CSEs are chosen for this PhD study due to their large impact. The first is a one-day event on 24th Jan 2016 that resulted in a lowest temperature of 3.3°C over HK, which is the ninth lowest on record and also the lowest temperature recorded since 1957 as reported by the HKO (HKO, 2016). The second is the one-day CSE of 15th Dec 2009 occurring over SCS and is a classical northerly winter monsoon surge, as described by Wu and Chan (1995). This event is associated with the passage of a cold front carrying cold air from a high-pressure region over Mongolia–Northern China to the SCS (Lim and Chang 1981). The third is a cold spell over 10th Jan to 5th Feb 2008. This CS that occurred from 10th January to 5th February 2008 induced extremely damaging ice storms, snow and frosts in south China (Zhou et al. 2009; Zhou et al. 2011; Yang et al. 2010; Lu et al. 2010) killed 129 people, damaged 11867 kilo/hectare crops and caused \$24 billion in economic losses (Zhao et al 2008; DCAS/NCC/CMA 2008). For HK, this cold spell lasted from 24th Jan to 10th Feb 2008. These three large scale strong CSEs are analysed at finer spatial and temporal resolutions using the high-resolution WRF model driven with the NCEP FNL dataset (NCEP, 2000). This WRF approach advances the earlier reported studies of CS using conventional meteorological data and satellite data. The WRF results are further compared with ECMWF reanalysed data and satellite-based ASCAT wind data. This thus represents a comprehensive comparative study in contrast to previous reported studies (e.g. Alpers et al. 2012; Hong and Li 2009) which are mostly limited to a single event or at a local scale and using reanalysis data and/or observed station data. It also enables a more comprehensive analysis of the convective mesoscale processes associated with the advent and dissipation of these CS leading to a more rigorous predictive criterion for forecasting CSEs at time scales of less than a week for South China and HK regions.

It is noted that the use of high resolution WRF modelling also helps reduce discrepancies and/or limitations in reanalyses data. Reanalyses are produced using different assimilation

techniques based on a variety of observational data (e.g., satellite, radar, surface stations, radiosonde) each having different temporal and spatial resolutions, and as such may have data discrepancies (Bao and Zhang, 2013) or limitations (e.g., Dee et al. 2011; Chelliah et al. 2011; Rood and Bosilovich 2009). The use of higher resolution numerical weather models improves the resolvability of mesoscale processes and reduces dependency on empirical physics parameterizations. A key benefit of such high-resolution models is their capability to simulate localised phenomena (e.g. Mass et al., 2014, Agustsson and Olafsson, 2014, Li et al., 2008). In the work here, we employed WRF version 3.7 for high resolution modelling over South China and HK regions.

4.2 Model Configuration & Experimental Design for South China and Hong Kong

Parameterization of physical processes is important in WRF modelling because small scale unresolved features can impact on the large resolved scales (Lorenz, 1969; Weeks, et al., 1997), leading to error growth, uncertainty and biases. The microphysics includes explicitly resolved water vapour, cloud, and precipitation processes while the cumulus parameterization schemes are responsible for the sub-grid scale effects of convective and/or shallow clouds. The surface layer schemes are responsible for calculating the exchange coefficients and friction velocities. Moisture fluxes and surface heat are computed by the LSM and serve as an input for the PBL scheme. The LSM scheme drives atmospheric processes providing the forcing from the ground to the lower part of the boundary layer which subsequently gets transported to the rest of the atmosphere and plays an important role in building soil temperature and moisture profiles and canopy properties. The PBL scheme determines the flux profiles within the well-mixed boundary layer and the stable layer, and thus provide atmospheric changes related to temperature, moisture (including clouds), and horizontal momentum in the entire atmospheric column. The physics options used in this PhD work have been discussed in Section 3.1.2 and illustrated in Figure 3.4.

As discussed in Section 3.2 under simulations, initial conditions for WRF simulations are prepared using the WRF Pre-processing System. This pre-processing system defines a physical domain and then interpolates the static fields (e.g., topography fields) to the domain grids. The processed data from the pre-processing system serve as input to the ARW real-data pre-processor, which then produces the lateral and initial boundary conditions for real data simulations. The real data simulations include vertical interpolation of meteorological fields to the WRF pressure levels (Skamarock et al., 2008). The WRF used is shown in Figure 4.1. It has a larger domain (D1) centered at the HKO Station ($22^{\circ} 18' 07''$ N; $114^{\circ} 10' 27''$ E) with 200×200 grid points and resolution of 9 km. A further nested smaller domain (D2) within D1 is also used as shown. It is also centered at HKO Station and has 151×151 grid points and resolution of 3 km. The nested grid D2 is used to resolve the bursts of low level cold air from the SMH approaching HK and spreading southward over the SCS during a CSE. Furthermore, this cold air interacts with the warm water of SCS and causes strong convective activity and generation of stratocumulus clouds (Johnson and Houze, 1987). Hence to resolve this, the atmospheric boundary layer is discretized with 31 uneven vertical levels with increased lower tropospheric resolution to capture the evolution of the low level cold air masses. Two-way nesting has been found to be generally superior to one-way nesting (Harris and Durran, 2010) and is also employed here.

Figure 5.1: WRF model domain D1 and its nested domain D2. Green shading represents land and blue the ocean. The domain is centered at HKO headquarter ($22^{\circ} 18' 07''$ N; $114^{\circ} 10' 27''$ E).

As discussed in Section 3.2 the WRF model is driven via initial and lateral boundary conditions using the NCEP FNL Operational Global Analysis data on $1.0^{\circ} \times 1.0^{\circ}$ grid resolution and 6 hr time intervals. The ECMWF ERA Interim data are used along with the wind fields from ASCAT, weather charts from HKO and meteorological information derived from HKO station data to validate the WRF simulated surface variables.

4.3 Model Evaluation & Case Studies

As mentioned earlier a CS onset is defined by HKO as (i) temperature drop $> 2^{\circ}\text{C}$ during the past 24 hr or (ii) mean temperature during the next 6 hours less than the 24- hourly mean temperature recorded 30 hr ago by $> 2^{\circ}\text{C}$. Furthermore, cold days are defined as days with minimum daily temperatures of 12°C or below. Since these CS at HK occur due to the cold air outbreaks via an extension of the SMH, we consider two key surface

variables, that of the Mean Sea Level Pressure (MSLP) and surface temperature (2m Temp used) for the examination of the CSEs. These two variables are used to demonstrate the WRF performance skill via a comparison with the HKO station data. The validation of our selected WRF physics and model resolution was performed by comparison with this HKO station data over 1st Jan -31st Jan 2016 with the station data being recorded at 00 UTC of each day. Here the WRF model was initialized on 1st Jan at 00 UTC using the NCEP FNL Operational Global Analysis data on $1.0^0 \times 1.0^0$ grid resolution. The model was driven via lateral boundary conditions from NCEP FNL at 6 hr time intervals. The model performed continuous simulations for 31 days on the same domain. For this WRF run, only the larger domain D1 was used due to the longer 31 days' simulation. This allowed the capturing of the circulation of the cold air masses as influenced by the terrain characteristics, and the interaction with the warm tropical water of SCS over an extended period. Furthermore, as this WRF run took 368 hr of computation time on a 16-core compute node consisting of 2 eight-core 2.4 GHz CPUs, the nested D2 was not exercised. However as shown below via performance statistics comprising mean errors, root mean square errors and correlation coefficient, the D1 grid resolution is sufficient.

A number of fluctuations are seen in the MSLP from the observed HKO data over this 31-day period (Figure 4.2a). The WRF simulations are able to capture daily variations of MSLP peaks though with a slight underestimation. The comparison of daily minimum temperature (Figure 4.2b) shows the WRF simulations (2m surface temperature) also being able to capture daily temperature trend along with the sharp rise/drop in the surface temperature. However, WRF shows an underestimation in simulating the minimum temperature. Mean errors, RMSE and correlation coefficient between the WRF results and station data, computed on daily intervals are shown in Table 4.1. The small mean errors and RMSEs and high correlation coefficients indicate that WRF simulations adequately captured the daily trend of MSLP and surface temperature.

While the above comparison shows that WRF model domain D1 is able to capture the variability in MSLP and minimum surface temperature at the daily level, subsequent

analysis of the three CSEs requires the finer nested domain D2 to better capture the sub daily spatial weather patterns over the specific days of the event.

Figure 4.2: (a) Daily mean MSLP and (b) Min surface temperature comparison between HKO Station (circles) and corresponding WRF simulations at 00 UTC of each day (triangles).

Table 4.1. WRF skill scores of MSLP and surface temperature at daily level over HKO for the period from 1st Jan -31st Jan 2016.

Parameters	Mean Errors	Root Mean Square Error	Correlation Coefficient
MSLP	0.708 hPa	0.898 hPa	0.994
Temp	1.125 $^{\circ}$ C	1.588 $^{\circ}$ C	0.961

4.3.1 Case Study: CSE 24th Jan 2016

The arrival of the CS in HK is marked by a sudden change in the synoptic conditions within 12 to 24 hours. This CS resulted in a lowest temperature of 3.3°C, the lowest recorded temperature since 1957. It originated from a strong high-pressure system with MSLP 1062 hPa seen over Mongolia (Figure 4.3a for the Weather Chart from HKO on 24th Jan 2016, 2:00 am HK local time). This synoptic pattern shows the establishment of a strong north-south pressure gradient between SH and SCS which only starts to weaken after 12 hr. This is seen in Figure 4.3b for 24th Jan 2016 2:00 pm HK local time where there is a weakened gradient across the mainland of China and north of HK. Starting from 21st Jan, the daily MSLP in HK started rising steeply from 1018.9 hPa reaching a max value of 1034.6 hPa on 24th Jan (Figure 4.4a). The daily minimum temperature also started dropping from 15.1 °C on 21st Jan to a minimum value of 3.1 °C on 24th Jan (Figure 4.4b) and the wind speed rose from 26.2 km/hr on 21st Jan to a max of 59.5 km/hr on 24th Jan (Figure 4.4c). Moreover, HKO recorded an all-time high monthly rainfall of 266.9 mm which is more than ten times the normal rainfall of 24.7 mm for the month of January. The day was reported to be the coldest day since the year 1957 with lowest temperature recorded at 3.3 °C versus the monthly averaged temperature of 16 °C. This monthly average temperature was a further 0.3 °C below the normal (monthly mean over 1981-2010) value of 16.3 °C (Table B.1, Appendix B). In addition to low temperatures the HKO recorded an all-time high monthly rainfall of 266.9 mm which is more than ten times the normal rainfall of 24.7 mm for the month of January (Table B.1, Appendix B).

At present HKO operational forecast includes predictions from Global Forecast System, European Centre for Medium Range Forecast, Japan Meteorological Agency and National Oceanic and Atmospheric Administration. However, for forecasting CSE 2016, all these models forecasted a minimum temperature of 10⁰ C on 24th Jan 2016 (Figure B.1, Appendix B) which was much higher than the actual observed temperature of 3.1⁰ C.

Figure 4.3: HKO Weather Chart for (a) 2 am on 24th Jan 2016 and (b) 2 pm on 24th Jan 2016

Source: http://envf.ust.hk/dataview/hko_wc/current/

Figure 4.4: Daily (a) mean MSLP, (b) min temperature and (c) mean wind speed from HKO Station (Lat:22° 18'07", Lon: 114° 10'27") over 1-31st Jan 2016 (horizontal axis indicates days from 1st Jan 2016).

The MSLP plot on 18 UTC on 23rd Jan 2016 (Figure 4.5a) as extracted from ECMWF shows a strong high-pressure system with MSLP of 1060 hPa over Mongolia and a strong

pressure gradient north of HK. The strong pressure gradient starts to weaken across mainland China and HK as it moves to the south of HK and towards the SCS as seen on 24th Jan 2016 at 06 UTC (Figure 4.5b). A comparison of the magnitudes and spatial distribution of the pressure peaks in the MSLP from ECMWF with the HK weather chart (Figures 4.5a and 4.3a and Figures 4.5b and 4.3b) shows spatial correlation indicating that the synoptic features are well established in the ERA interim datasets. The 10m wind field extracted from ECMWF, ERA Interim also shows northeasterly winds with speed in the range 25-30 km/hr approaching HK from mainland China (Figure 4.5c) on 23rd Jan 2016 18 UTC. On 24th Jan 2016 06 UTC, a continuous and widespread band of wind from Siberian-Mongolian region is seen traveling towards mainland China and into HK (Figure 4.5d). During this 12 hours, northeasterly winds with strength of 40-50 km/hr are seen over HK which further increases in strength as it moves from south of HK towards the SCS. The 2m temperature field from ECMWF, ERA Interim also shows freezing temperature prevailing over the over the entire region of Mongolia and south-central China on 23rd Jan 2016 18 UTC (Figure 4.5e). On 24th Jan 2016 at 06 UTC there is a widespread and continuous band of cold air crossing the SCS (Figure 4.5d). Along with this there is a rise in the surface temperature over the SCS (Figure 4.5f) due to the interaction of these cold air masses with the warmer SST leading to unstable near surface atmospheric boundary layer. The air temperature north of HK increases in the range from 3- 6 °C to 9 - 12 °C after the passage of this band of cold air.

The near surface wind fields on 24th Jan 2016 12 UTC from ASCAT are shown in Figure 4.6. There is a distinct band of high northeasterly wind with speed 50-60 km/hr (yellowish colour) adjacent to the coast at HK near the SCS (Figure 4.6a) with these winds having a maximum speed of 75 km/hr (amber colour). These strong winds over the SCS then heads towards Malaysia maintaining the north easterly flow direction (Figure 4.6b). The ECMWF wind data (Figure 4.5d) shows similar strength winds in the SCS. However due to the limitations of ASCAT 25 km horizontal resolution, the wind field close to the coast of HK could not be analysed. This is because land targets contaminate the backscattered radar signal at resolution distances smaller than the ASCAT cell resolution.

Figure 4.5: ECMWF, ERA Interim plot for MSLP ((a) & (b)), 10m Wind ((c) & (d)) and 2m temperature ((e) & (f)) on 18 UTC 23rd Jan 2016 (left panel) and 06 UTC 24th Jan 2016 (right panel). The red box indicates the study region. In figure (c) and (d) arrows represents the wind directions.

Figure 4.6: ASCAT data for (a) 10 m wind speed and (b) 10 m wind direction for 12 UTC 24th Jan 2016. The HK and the nearby SCS and coastal regions are encircled.

Figure 4.7a shows the hourly MSLP values recorded at the HKO Station on 24th Jan 2016 at HK local time. Starting with a MSLP of 1031 hPa at 1 am, the MSLP reached to a maximum value of 1037.7 hPa between 10-11 am. With the rise in MSLP, the lowest temperature of 3.3 °C (Figure 4.7b) was recorded on 24th Jan 2016 6 am. This MSLP value is also considered to be highest MSLP value recorded till the date (HKO, 2016).

Figures 4.7c to g show the corresponding WRF simulations. The WRF model ran with the nested D2 domain over 23rd Jan 2016 12 UTC to 24th Jan 2016 12 UTC. The WRF results show a sudden and consistent increase in MSLP from 2 am to 9 am (Figure 4.7c). The

increasing MSLP climbs to a maximum value of 1034.81 hPa at 11 am after which it then starts to decrease. Furthermore, WRF predicts a continuous drop in temperature from 8 °C at 2 am, reaching a minimum of 3.52 °C at 6 am (Figure 4.7d). The model also predicts the sudden increase in wind speed from 2 am reaching a max speed of 42 km/hr at 6 am (Figure 4.7e), as associated with sudden drop in temperature and steep increase in MSLP. The wind speed then starts dropping with the increase of air temperature. The hourly WRF simulation at 01 UTC (9 am) on 23rd Jan shows a wide spatial variation of MSLP over HK (Figure 4.7f). It also shows that during the outbreak of the strong surge the regions situated north west region of HK are likely to come under the influence of highest MSLP of 1036 hPa (brown shading in Figure 4.7f) with wind speed of 45 km/hr. Furthermore, cold air at 4 °C (Figure 4.7g) and speed in the range 50-60 km/hr interact with the warm water with SST at 24°C (Figure 4.7h). This large temperature difference leads the near surface boundary layer becoming unstable and triggering atmospheric convection along with high intensification in momentum transfer (wind stress) from the marine boundary layer towards the sea surface. This further increased the near surface wind due to the upward momentum transfer from surface waves sufficiently exceeding the downward turbulent momentum flux, resulting in acceleration of near surface wind leading to strong CS in HK and the nearby region (Figure 4.7g for the surface temperatures). Overall the WRF surface variables with 12 hour of lead time thus successfully captured the sudden strengthening, advancement and later gradual weakening of the Jan 24th CSE.

Figure 4.7: HKO Station hourly plots of (a) MSLP and (b) temperature from 02 am to 02 pm local HK time on 24th Jan 2016 (indicated by black box). WRF-D2 time series plots for (c) MSLP, (d) 2m temperature and (e) 10m wind from 2 am 23rd Jan to 2 pm 24th Jan 2016 (horizontal axis indicates hour from 02 am on 24th Jan to 02 pm local HK time 24th Jan 2016). WRF-D2 spatial plots for HK region (f) MSLP (g) 2 m temp at 9 am (01 UTC) on 24th Jan 2016. WRF-D1 spatial plots for South China (h) 10 m wind & SST contour at 9 am on 24th Jan 2016. In Figure f and g, HK region is represented in red colour while in Figure h, HK is represented by a blue dot and isotherms are represented by black lines. In Figures f, g, and h the coastal boundaries are represented by black thick line.

Source for figure 4.7a & b: http://envf.ust.hk/dataview/hko_wc/current/

4.3.2 Case Study: CSE & Cold Spell Event 24th Jan – 16th Feb 2008

HK experienced its longest persisting cold spell in 40 years over the period from 24th Jan to 16th Feb 2008 (HKO, 2016). This surge event is considered as a major forecasting failure (Zhou et al., 2011). The MSLP plot prepared from ECMWF, ERA interim data shows a strong high-pressure system (1050 hPa) over Mongolia and a strong pressure gradient ahead of HK on 24th Jan 2008 00 UTC as indicated with a red box (Figure 4.8a). The daily MSLP values from HKO Station shows an overall increasing trend computed using Mann-Kendall (MK) test with Z statistics of 2.81 and p value of 0.003 during this period with the peak value of MSLP rising to 1023.1 on 14th Feb (Figure 4.9a). The daily minimum temperature shows an overall decreasing trend using MK test with Z statistics of (-)1.62 and p value of 0.116 dropping to a minimum value of 7.9 °C on 3rd Feb (Figure 4.9b). Moreover, the temperature during this period continually remains below 12°C (Figure 4.9b), which is 2°C below the normal for Jan and Feb (Table B.2 & B.3, Appendix B) versus the Jan monthly averaged temperature of 15.9 °C. This Jan monthly average temperature was a further 0.2 °C below the normal value (monthly mean over 1971-2000) of 16.1 °C. The total monthly rainfall was 33.3 mm which was higher than the normal rainfall of 24.9 mm for January (Table B.2, Appendix B).

There was also an overall increase in wind speed (Figure 4.9c), implying intrusion of cold air masses over HK during the entire period. Under the combined effect of the NEM and a broad rain band, cold and rainy weather prevailed in the ensuing days with highest rainfall of 19.2 mm on 25th Jan (Figure 4.9d). The MSLP chart on 16th Feb 00 UTC (Figure 4.8b) shows a weakened pressure gradient across mainland China into HK.

Figure 4.8: ECMWF, ERA plot chart for MSLP for (a) 00 UTC 24th Jan 2008 and (b) 00 UTC 16th Feb 2008. The red box indicates the study region.

Figure 4.9: Daily (a) mean MSLP, (b) min temperature, (c) mean wind speed and (d) total rainfall from HKO Station over 24th Jan-16th Feb 2008 (Horizontal axis indicates days from 24th Jan to 16th Feb 2008). Dotted lines in (a) to (b) indicates fitted linear trends.

The WRF model run with the domain D1 was initialised on 24th Jan 2008 00 UTC. Figure 4.10a shows the daily MSLP values at 8 am from the WRF simulations over 24th Jan-16th Feb 2008. The max MSLP during the 24 days WRF simulation occurred on 14th Feb which is in agreement with the observed data from HKO (see Figure 4.9a, Figure 4.10a, and Figure B.2 Appendix B) with an overall increasing trend computed using Mann-Kendall test with Z statistics of 2.90 and p value of 0.009. Furthermore, WRF also simulated an overall decreasing trend with Z statistics of (-)1.36 and p value of 0.153 in the daily minimum temperature reported at 8 am. The minimum temperature during the 24 days WRF simulation occurred on 3rd Feb 8 am, which is also in agreement with the observed data from HKO (see Figure 4.9b and Figure 4.10b). The MSLP values rose continuously from 5th to 14th Feb with the exception of 10th and 11th Feb (Figure 4.10a). In addition, there was a sudden drop in surface temperature of 7.67 °C over 5th to 6th Feb and the temperature remained below 11.5⁰ C till 15th Feb (Figure 4.10b). Hence, we infer that the cold spell at HK during this period was mostly maintained by the arrival of CS.

Conversely the earlier period from 24th Jan 12 UTC to 4th Feb 12 UTC had the cold spell mostly driven by strong winds flowing through the Straits of Taiwan towards HK (Figure 4.10c for WRF 10m wind at 8 am, 25th Jan), bringing enough moisture from the ocean resulting in daily rainfall of 19.2 mm on 25th Jan (Figure 4.9d). In addition, on 1st Feb 00 UTC there is an anticyclone residing over south eastern China (Figure 4.10d) that remains stationary till 3rd Feb 18 UTC. This resulted in an abrupt increase in north easterly cold air travelling inland along the east coast of China, reaching HK as simulated using WRF (Figure 4.10e) and led to the HK daily minimum temperature dropping to a minimum of 7.9⁰C on 3rd Feb. The longest persisting cold spell in HK therefore arose from two different episodes of advection of winds towards HK. The first period from 24th Jan to 4th Feb was mostly dominated by strong easterly winds through the Strait of Taiwan that resulted in heavy to moderate rainfall over HK. The second period of 5th Feb to 16th Feb had cold air arriving from Mongolia towards mainland China as a CS reaching HK as north easterly winds. This is due to strengthening of the pressure gradient between

Mongolia and SCS. This led to the sudden drop in temperature, increased MSLP and overall increase in the wind speed.

Figure 4.10: WRF-D1 (a) MSLP and (b) temperature from 24th Jan to 16th Feb 2008 at 00 UTC; (c) WRF 10m wind at 8 am on 25th Jan; (d) HKO Weather chart for 2 pm on 2nd Feb 2008; (e) WRF 10m wind at 8 am on 3rd Feb 2008. Dotted lines in (a) to (c) indicates fitted linear trends.

4.3.3 Case Study: CSE & Cold Front Event 15th Dec 2009

A strong north-south pressure gradient developed over this period and a cold front crossed the coast of HK and moved in southerly direction over the SCS on 15th Dec 2009. Figure 4.11 shows the north-south pressure gradient and the cold front ahead of HK at 00 UTC 15th Dec 2009 (Figure 4.11a) and later at 00 UTC 16th Dec 2009 (Figure 4.11b). During this period, the monthly averaged temperature remained 4.3 °C below the normal value (monthly mean over 1971-2000) of 15.6 °C (Table B.4, Appendix B). The daily mean temperature remained 4.4 °C below the normal temperature of 17.7 °C. This event is a classical northerly winter monsoon surge or cold air outbreak event described in Wu and Chan (1995) and is characterized by sharp drop in temperature and sudden increase in wind speed.

Figure 4.12a shows the hourly MSLP values recorded at HKO Waglan station on 15th Dec 2009. The MSLP started to rise from 5 pm and reached its peak of about 1019 hPa at 11 am. As reported by HKO, the cold front crossed Waglan Island between 1200 and 1300 UTC. After the passage of the cold front, the temperature at Waglan station dropped from 21 °C to 13 °C. WRF with the nested D2 domain was ran from 15th Dec 2009 00 UTC to 16th Dec 2009 00 UTC. The WRF results showed a sudden and consistent increase in MSLP from 5 pm that reached a peak value of 1012.7 hPa at 11 pm (Figure 4.12b). Furthermore, the model predicted a steep drop of 4.71 °C in temperature between 6 pm to 11 pm (Figure 4.12c), with a sudden and rapid increase in wind speed from 3.3 km/hr to 51.8 km/hr over 5 pm to 7pm (Figure 4.12d). The wind strength then decreased and levelled off to 42.5 km/hr at 11 pm. The 10-m wind field from WRF at 06 UTC 15th Dec 2009 showed the near-surface wind in the range of 10-15 km/hr and the temperature in the range of 20-22 °C over Waglan Island (Figure 4.12 c & d). On 13 UTC of the same day, parallel to the coastline, a broad band of increased wind speed is seen in WRF results near HK and adjoining coastal regions (Figure 4.12e). The WRF model therefore distinctly showed the boundary of the high and low wind speed areas, which formed the wind front. Behind this front, there was strong advection induced by the large air-sea temperature

difference which generated clouds and resulted in significant precipitation (HKO, 2009). The above analysis of the time series and spatial maps from the WRF results showed that the front was associated with increased MSLP, sudden drop in temperature and steep increased in wind speed that would have crossed Waglan Island between 12 to 13 UTC on 15th Dec 2009.

Figure 4.11: HKO Weather chart for (a) 00 UTC on 15th Dec 2009 and (b) 00 UTC on 16th Dec 2009.

Figure 4.12: (a) HKO Waglan station MSLP from 02 pm to 11 pm HK local time on 15th Dec 2009 (indicated by black box); WRF-D2 (b) MSLP; (c) 2m temperature; (d) 10m wind from 2pm 15th Dec 2009 (horizontal axis indicates hour from 02 pm 15th Dec to 00

am 16th Dec local HK time); (e) WRF-D1 10 m wind and 2m Temp for 13 UTC 15th Dec 2009. The red dot in Figure e represents HK region.

Additional WRF model diagnostics are performed to investigate the role of the Planetary Boundary Layer Height during the occurrence of CSs. The analysis is performed for CSE 2016, CSE 2009 and CSE 2008 events. A particular behavior that of PBLH increasing is observed during the onset of these three surges, suggesting a possible role of PBLH. However, further investigations are required to define its particular role. The details of the results and discussions are presented in Appendix C.

4.4 Design of new operational criteria for forecasting CS over South China and HK

The analysis of the three unique CSEs over HK indicated that the advancement of the CS from the SMH towards SCS is dominated by a strong SMH and also a pressure gradient north of HK. To quantify the SMH high, we first set the domain of the SHM as guided by previous studies though it is noted there is no common consensus. For example, domain of SMH between 40–65 °N and 80–120 °E was defined by Panagiotopoulos et al. (2005), while in the previous study by Gong and Ho (2001), they restricted the northern boundary of SMH to 60 °N but extended its western flank to 70 °E. Woo et al. (2012) considered the domain to be over 90–115°E and 35–55°N. Here we adopted the overlapped area from these reported domains, i.e. 40°N-60°N and 80°E-120°E, for analysis of the SMH. This defined domain captured well the geographical location of the SMH intensity during Dec 2009, Jan & Feb 2008 and Jan 2016 as shown in Figure 4.13 for the monthly maximum computed from the daily max MSLP values over 00, 06, 12 and 18 UTC. This domain definition has also been previously used also by Wu and Wang (2002), Jhun and Lee (2004) and Wu et al. (2006) and for studying the intensity of SMH.

The cold air outbreaks from SMH moving towards the tropical latitudes and eastern longitudes are driven by strengthening of the pressure gradient between the SMH and the SCS (SCS domain defined here as 115 °E -120 °E and 15 °N -20 °N) and between the

SMH and Low Pressure Systems (LPS) developed within the domain of $145^{\circ}\text{E} - 180^{\circ}\text{E}$ and $30^{\circ}\text{N} - 55^{\circ}\text{N}$, i.e. between the eastern coast of Japan and the Pacific region to be discussed later. Here the pressure gradients are computed based on difference between the centres of the max MSLP and min MSLP and the distances between them using the ‘haversine’ formula, and using the ECMWF, ERA Interim analysis field at 00, 06, 12 & 18 UTC. It is noted that while a strong SMH MSLP value of 1067.9 hPa on 22nd Jan 2016 impacted HK as an intense CSE due to a strong SMH-SCS pressure gradient being established, a slightly lower SMH MSLP value of 1063.1 hPa on 12th Jan 2008 did not impact the HK region even as a lower intensity CSE. This was because the SMH-SCS pressure gradient was not well established. Table 4.2 shows that a SMH-SCS pressure gradient of 1.165 Pa/km being established at 06 UTC on 23rd Jan 2016 for the Jan 2016 CSE. The Jan-Feb 2008 cold spell (5th to 14th Feb) was driven by the SMH-SCS pressure gradient with a pressure gradient of 0.669 Pa/km (Table 4.2) established on 4th Feb 2008 06 UTC i.e. 18 hr before the arrival of the CS on 5th Feb 2008. It is further noted that the max SHM high of 1051.1 hPa for Feb 2008 was on 11th Feb (Table 4.3) i.e. well after the CS arrival. Similarly, for the Dec 2009 event, the month’s max SMH MSLP was 1059.2 hPa on 16th Dec 2009 which was after the CS arrival in HK on 15th Dec. Rather the surge was driven by the SMH-SCS pressure gradient of 0.87 Pa/km, established on 14th Dec at 06 UTC (Table 4.2) i.e. 18 hr before the arrival of the CS. The above indicates that the establishment of SMH-SCS pressure gradient with a lead time of 18 hr, rather than the SMH high alone, is a reliable indicator of HK CSEs.

Figure 4.13: ECMWF, ERA Interim plot for max MSLP (a) Jan 2016 (b) Dec 2009 (c) Jan 2008 and (c) Feb 2008 computed from daily maximum values over 00 UTC, 06 UTC,

12 UTC, and 18 UTC. Coastal boundaries are represented by black thick line. The black box indicates the defined SMH domain within 40°N-60°N and 80°E-120°E.

Equally importantly we also observed the formation of a series of intense to very intense short-lived Low Pressure System (LPS) located near the western coast of Japan and eastern pacific within the domain 145°E - 180°E and 30°N - 55°N with central pressure values varying between 953 to 1000 hPa. These LPS persists for 18-24 hr and which significantly influences the pressure gradient developing between SMH-SCS. The LPS were most evident for the Jan-Feb 2008 event, with such LPS occurring over 13th-14th Jan, 24th-27th Jan, 29th-30th Jan, 3rd-4th Feb, 7th-8th Feb and 13th-14th Feb (Figure 4.14). As tabulated in Table 4.2, the SMH-LPS pressure gradient for Jan 2016 CSE starts to weaken continuously from 1.62 to 1.46 Pa/km on 23rd Jan over 06 UTC to 18 UTC. Furthermore, during the same period, the pressure gradient between SMH-SCS strengthened from 1.17 to 1.51 Pa/km. Similarly, for Feb 2008 and Dec 2009 CSEs the SMH-LPS pressure gradient continuously weakened over the 18 hr shown (Table 4.2) while the SMH-SCS pressure gradient strengthened. This development of the two pressure gradients between SMH-LPS and SMS-SCS led to the formation of two major cold wind tracks, one moving towards the SCS (black arrow, Figure 4.15a, b &c) and other towards the LPS (white arrow, Figure 4.15a, b &c). The Jan part of the HK Jan-Feb 2008 cold spell event had the event's strongest SMH MSLP of 1063.1 hPa occurring on 12th Jan 2008 06 UTC (Table 4.3). However, the SMH-SCS pressure gradient weakened while the SMH-LPS pressure gradient strengthened over 13th Jan (Table 4.2). Furthermore, the SMH-LPS pressure gradient was significantly larger than the corresponding gradient values during the CSEs. This affirms that the SMH alone is a poor indicator of HK CSEs and that the SMH-LPS determines the (non-)onset of such events. Similar observations hold for the 17th Dec 2009 which is a non-CSE (Table 4.2). While the months maximum SMH occurred on the 16th, the SMH-SCS pressure gradient remains approximately constant on the 17th while the SMH-LPS pressure gradient had a much larger value, again in comparison to the CSE values.

Table 4.2. Pressure Gradient between SMH and SCS, and between SMH and LPS from ECMWF, ERA Interim for HK CSEs (non-shaded cells) and HK non-CSEs (shaded cells).

(Non-)CSE	Date	Pressure gradient (Pa/km): SMH and SCS	Pressure Gradient (Pa/km): SMH and LPS
Jan 2016	23rd at 06 UTC	1.165	1.616
	23rd at 12 UTC	1.228	1.503
	23rd at 18 UTC	1.514	1.462
Dec 2009	14th at 06 UTC	0.874	1.669
	14th at 12 UTC	0.902	1.657
	14th at 18 UTC	0.947	1.626
Dec 2009	17th at 12 UTC	0.887	2.295
	17th at 18 UTC	0.984	2.286
	18th at 00 UTC	0.897	2.213
Jan 2008	13th at 12 UTC	1.024	2.040
	13th at 18 UTC	1.008	2.475
	14th at 00 UTC	0.942	2.650
Feb 2008	4th at 06 UTC	0.669	1.650
	4th at 12 UTC	0.721	1.565
	4th at 18 UTC	0.829	1.561

Table 4.3. Max MSLP Values from ECMWF, ERA Interim within the SMH domain (40°N -60°N to 80°E-120°E for Jan 2016, Dec 2009 and Jan-Feb-2008.

Year	Date & UTC	Max_MSLP (hPa)	Centre of Max MSLP (Lat, Lon)
2016	22 nd Jan at 00 UTC	1067.90	51.0°N, 93.75°E
2009	16 th Dec at 06 UTC	1059.20	60.0 °N, 111.0°E
2008	12 th Jan at 06 UTC	1063.10	57.75 °N, 108.0°E
	11 th Feb at 06 UTC	1051.07	46.5 °N, 87.75°E

Figure 4.14: ECMWF, ERA Interim plot for MSLP (a) on 13th Jan 2008 00 UTC; (b) on 24th Jan 2008 00 UTC; (c) on 29th Jan 2008 00 UTC; (d) on 3rd Feb 2008 00 UTC; (e) on 7th Feb 2008 00 UTC; and (f) on 13th Feb 2008 00 UTC showing formation of frequent LPS during Jan-Feb 2008 between 140°E - 180°E and 30°N - 55°N. The coastal boundaries are represented by black thick line and HK region with a red dot.

Taken together, the three CSEs cases analysed indicate a threshold SMH-SCS pressure gradient value of 0.74 Pa/km, maintained for at least 18 hr, and subjected to SMH-LPS pressure gradient value not exceeding 1.59 Pa/km for occurrence of CSEs in HK. This is based on the 18-hr averaged values for 4th Feb 2008 event which had the smallest value SMH-SCS gradient. The severity of the CS may further be assessed based on the strength of the SMH-SCS pressure gradient during the 18 hours before the CSE. The above cases also demonstrate that the SMH-LPS pressure gradient can lead to the bifurcation of the cold air mass from the SMH and thereby affect the evolution of the CS event. This is shown in Figure 4.15a, b & c for the wind fields during the three CSEs. The right track (white arrow) is dominated by the pressure gradient towards LPS, with winds moving through Yellow Sea and extends towards the eastern Pacific. The left track (black arrow) dominated by the pressure gradient towards SCS with winds moving through mainland China towards HK and extending to SCS. Thus, the cold air outbreak originating at SMH as driven along these tracks extends towards the tropical latitudes or the eastern longitudes. For instance, the cold air mass driven towards the SCS during the Jan 2016 event also strongly impacted tropical latitudes such as the regions of Vietnam and Taiwan. This is readily quantified via the WRF results (see Figure 4.16 for hourly 2m temperatures). There was a steep drop in temperature in Nantau, Taiwan from 10.53 °C to 6.13 °C within one hour, while the temperature dropped from 3.28 °C to 0.53 °C in Dong Dang, Vietnam within six hours. The same event also had cold air mass driven towards the LPS that impacted strongly on eastern longitudinal regions such as the northeast regions of China, southern coast of South Korea and southwestern coast of Japan. This is quantified via the ECMWF data (see Figure 4.17 for six hourly 2m temperatures). The temperature dropped suddenly from 7.06°C to 0.204°C in Jinan, China while in Qingdao the temperature dropped continuously from -8.3°C to -12.6°C. The temperature similarly dropped steeply from -3.1°C to -7.9°C and from 3.5°C to -2.04°C in Busan, South Korea and in Fukuoka, Japan, respectively.

Figure 4.15: ECMWF, ERA Interim plot for 925 hPa Wind (a) on 24th Jan 2016 06 UTC; (b) on 15th Dec 2009 12 UTC; and (c) on 5th Feb 2008 12 UTC showing formation of LPS within the domain 140°E - 180°E and 30°N - 55°N. The white thick arrow indicates winds moving towards the LPS and the black thick arrow winds moving towards southern China. Wind directions are represented by small black arrows, coastal boundaries by black thick line, HK region with a red dot.

Figure 4.16: WRF hourly 2m temperature for Nantau (triangle), HKO (square) and Dong Dang (circle) from 18 UTC 23rd Jan to 00 UTC 24th Jan 2016 (horizontal axis indicates hour from 18 UTC 23rd Jan 2016).

Figure 4.17: ECMWF six hourly 2m temperature for Jinan (triangle), Fukuoka (square), Busan (circle) and Quingdao (diamond) over 00 UTC 23rd Jan to 00 UTC 24th Jan 2016 (horizontal axis indicates hour from 18 UTC 23rd Jan 2016).

4.5 Summary

CS over South China and HK have been studied using a combined approach of high resolution WRF-ARW model, Reanalysis data from ECMWF & NCEP, satellite data from ASCAT and station data from HKO, this being one of the very few such comprehensive studies. Specifically, we examined the three important and unique South China/ HK CS viz. an intense surge of Jan 2016, a long persisting cold spell over Jan-Feb 2008 and a classical northerly surge with cold front in Dec 2009. Under the combined effect of the pressure gradients developed between SMH-SCS and SMH-LPS, it is shown that the CS from the SMH bifurcates into two major tracks, one moving towards the SCS and the other towards the LPS. It is noted that while a strong MSLP magnitude at the SMH is necessary for a strong cold air outbreak, a better and reliable indicator of HK CS is the pressure gradient between the SHM and SCS. The three cases studied indicates a CS onset towards HK is associated with a SMH-SCS pressure gradient above a threshold value of 0.78 Pa/km over 18 hr duration and subjected to a SMH-LPS pressure gradient not exceeding 1.59 Pa/km. However, more HK CS cases should be examined to confirm this result and to fine-tune the threshold values.

The short duration WRF model simulations also qualitatively and quantitatively reproduced the sudden increase in MSLP, sharp drop in temperature and increased wind speed generated by the three CSEs for South China, HK region. However, to better predict the frontal LPS and associated weather features during CSEs, it is most important to investigate the mechanism that leads to these CSEs in East Asia and towards SEA. Such an investigation requires a detailed understanding of the complex dynamic mechanisms of CS and associated atmospheric circulation patterns to reasonably estimate the occurrence and intensity of CSEs. Such an analysis is conducted in Chapter 5.

Chapter 5

5. Large scale pattern of CSE Over East Asia

The potentially huge socioeconomic impacts caused by CSEs lead to important questions about the formation of these extreme events on a large atmospheric scale, and how much predictability existing atmospheric models can provide. The latter is in past driven by the poor representation and understanding of dynamical processes in atmospheric models that substantially contribute to large uncertainties in CSE prediction. In this chapter we investigate the mechanism and define a new theoretical framework that links the CSE to the large-scale circulation patterns in the atmosphere, specially the SMH, AL and JS. This framework further allows as assessment of the likelihood of CSEs given the monthly behavior of these atmospheric patterns.

5.1 Jet stream

The JS over East Asia is an important large-scale circulation system of the atmosphere that significantly influences weather and climate over the Asian-Pacific region. Previous research has shown that the JS over East Asia is associated with synoptic-scale phenomena such as monsoon, storm track activity, cyclogenesis, frontogenesis, and blocking (e.g., Bell et al. 2000; Gao and Tao 1991; Dole and Black 1990; Kung and Chan 1981; Zeng 1979; Palmen and Newton 1969). During November to March, strong temperature gradients form between the SMH and the warm air masses to the south and east which can lead to wintertime East Asian CSE that exert tremendous economic and societal impacts on the countries in East Asia. Each winter, approximately 10 CSEs occurs across East Asia (Chen et al. 2004). Their onset often causes snowfall and freezing precipitation over east Asian regions (Jeong et al. 2008; Chen 2002; Ding 1994; Boyle and Chen 1987) and are the root causes of severe weather over the SCS (Chan and Li 2004). Zhang and Xiao (2013) found that the intensity of the JS is intimately related to the atmospheric general circulation in mid-latitude East Asian region and significantly

correlated with the EAWM. Furthermore, recent research on JS suggests an association with the cold and warm air activities during the boreal winter and the Mei-Yu season during the summer (Huang et al. 2014; Ye and Zhang 2014; Li and Zhang 2014; Zhang and Xiao 2013; Liao and Zhang 2013). The warming in the Arctic may be intensifying the effects of the jet stream's position, which in the winter can cause extreme cold weather, such as the winter of 2014/15 which saw record snowfall levels in New York (Overland et al. 2016). However, despite this possible linkage of the JS with the outbreak of CSE, the relationship between JS and CSE occurrence in East Asian regions is not fully explained and which is the focus of this chapter.

5.2 Siberian Mongolian High and Aleutian Low Systems

During the northern-hemisphere wintertime condition, monsoonal northerlies prevail over the region between the SMH over the Eurasian continent and the Aleutian Low (AL) over the North Pacific, bringing cold air into the midlatitude. The climate of East Asia during the winter months being dominated by EAWM (e.g., Chang et al., 2006) is closely associated with the development of a cold core high pressure system over the SMH region that brings the monsoon from the land mass towards South China (Figure 5.1).

Figure 5.1. Schematic of EAWM progressing towards South China, HK. Cold Core over SMH is indicated by 'COLD' in the figure.

Source: <http://www.hko.gov.hk>

Semi-permanent highs and lows such as the SMH and AL are the persistent pressure systems that appear over an area during the year and affect the weather and climate systems over a large area. The global semi-permanent systems for the month of January are shown in figure 5.2. Large contrast in the position of the pressure systems are observed between the months of January and July. The SMH is a semi-permanent pressure system during the winter and is a cold core, high-pressure system with central pressure reaching more than 1035 hPa. It dominates much of the Eurasian continent and is maintained by large-scale subsidence of air masses and strong radiative cooling (Ding and Krishnamurti 1987). With this cold core, high pressure system, northwesterly winds during the EAWM is intensified along the eastern flank of the SMH that brings severe CSE and heavy snowfall events to East Asia (Li and Yang, 2010; Jhun and Lee, 2004). The amplification of the SMH is recognized as an essential factor for the generation and maintenance of CS (Takaya and Nakamura 2005a; Gong and Ho 2002; Zhang et al. 1997).

Another semi-permanent pressure system is located over the Aleutian Islands known as AL. This system represents one of the main climatological drivers playing a vital role in the sea–air interaction for the North Pacific. It is one of most intense (lowest) pressure system during the winter but practically disappears during the summer (Overland et al., 1999). Previous studies have found that variability of the AL can even influence climate anomalies in remote regions (Zhu and Wang, 2010; Ye and Duan, 2015). Further research indicated that the northerly surge greatly affects not only convective activity over Asia but also rainfall distributions in the tropics near the equator and over the Indonesian Maritime Continent in the Southern Hemisphere (Hattori et al. 2011; Wu et al. 2013; Pullen et al. 2015). Up to now, it has been difficult to fully show the specific propagation processes of the CSE from the statistical studies due to the limited knowledge on the underlying mechanism behind the correlative relationships. Previous research on CSE are further mostly limited to a typical case study or a local scale CSE (as discussed in detail in Section 4.1) and more recently by using satellite data, in particular synthetic aperture radar data (Alpers et al. 2012, 2015). A few studies have also identified strong CSE effects on spatiotemporal variations in air quality on a large scale (e.g. Qian, 2009; Wang, 2016). A recent study indicates that the rapid strengthening of SMH is linked to the phase transition of AO that favors southward intrusion of cold air towards East Asia (Geng, 2017). However, despite the above-mentioned works, little attention has been paid to examining the potential mechanism guiding the outbreak of cold air masses from SMH, and effects of AL on the surge propagation to lower latitudes. This is explained in the following Section 5.3 to 5.4 of this PhD thesis.

Figure 5.2. Semi-permanent pressure systems during Jan. Low-pressure systems (L) are shaded in blue and purple. High-pressure systems (H) are in red and orange.

Source: <https://www.meted.ucar.edu>

5.3 Possible Mechanism on the outbreak of CSE

During the EAWM, formation of northerly CS take place when cold air is advected towards south along the eastern edge of the SMH. The CSE can reach lower latitudes where it manifests as strengthened northerly or northeasterly winds. CSE over East Asia are characterized by an abrupt drop of surface air temperature and an expansion of the SMH. Based on these, previous literatures have proposed criteria for identifying CSE (Jeong and Ho 2005; Chen et al. 2002, 2004; Zhang et al. 1997). Depending on the thermodynamic characteristics of the pre-incursion environment, CSE may also have a substantial impact on convection and rainfall in the tropics. In this chapter, we consider the large-scale effect on the three CSEs that occurred in 2008, 2009 and 2016. To

establish the relationship between cold air outbreaks in winter and its progression towards lower latitudes, we analyzed these unique CSEs using Reanalysis data sets from ECMWF, ERA Interim. The domain of SMH is defined based on the earlier definition in Section 4.4. Figure 5.3 represents the monthly averaged MSLP derived from the daily max MSLP values of ECMWF, ERA Interim, over 00, 06, 12 and 18 UTC for the month of Jan 2016, Dec 2009 and Jan & Feb 2008. During Jan 2016, the SMH is dominated by intense MSLP values varying from 1025 hPa to 1055 hPa at daily level suggesting possible extremely cold and dense air over the region. The strongest monthly MSLP attained in Jan 2016 (1064 hPa) is next followed by that in Jan 2008 (1060 hPa), Dec 2009 (1058 hPa) and Feb 2008 (1050 hPa) respectively.

As mentioned earlier the AL is an important indicator of winter climate as demonstrated by previous studies (e.g., Parkinson, 1990; Trenberth and Hurrell, 1994; Niebauer et al., 1999). It has also been found that the variability of the AL can significantly influence climate anomalies over remote regions (e.g., Zhu and Wang, 2010; Yu and Kim, 2011; Park et al., 2012; Xiao et al., 2014). To quantify the intensity of low pressure systems occurring during the winter season, we propose the domain of AL as between 25–75 N, and 160 E–130 W (Figure 5.4). Figure 5.4 shows the monthly averaged MSLP derived from the daily min MSLP values of ECMWF, ERA Interim, over 00, 06, 12 and 18 UTC for the month of Jan 2016, Dec 2009 and Jan & Feb 2008. During Jan 2016, the AL is dominated by intense MSLP varying from a maximum of 966 hPa and dipping to minimum value below 950 hPa indicating deep convection over the region. The corresponding values are Feb 2008 (950 hPa), Dec 2009 (954 hPa) and Jan 2008 (958 hPa). The intense monthly MSLP in Jan 2016 is followed by that in Feb 2008 (950 hPa), Dec 2009 (954 hPa) and Jan 2008 (958 hPa) respectively.

We believe that the physical process of the upper troposphere need to be considered as both the lower and upper troposphere are likely related when explaining the mechanisms for CS progression. As discussed in Sections 2.1 & 5.1 the meandering JS bands of fast winds are driving factors for weather in the midlatitudes. We postulate that JS being an

important feature of upper troposphere, always plays a key role in the occurrence of synoptic disturbances at the lower troposphere originating from SMH and AL. Though it has also been reported that JS provides favorable conditions for the occurrence of CSs in East Asia (Gong and Ho 2004; Jeong and Ho 2005). To the best of the authors knowledge, this is the first study to analyse the possible link of JS along with SMH and AL in CS outbreaks for East Asia. The JS over East Asia is associated closely with many synoptic-scale phenomena such as cyclogenesis, frontogenesis, monsoon, blocking, and storm track activity (e.g., Tao 1991; Bell et al. 2000). Zhang and Xiao (2013) found that the intensity of the JS is intimately related to the atmospheric general circulation in mid-latitude East Asian region and significantly correlated with the EAWM. During the EAWM, the strong westerly JS dominates the circulation system in the upper troposphere over East Asia. Previous literatures indicated that the CS during the EAWM are often preceded by upstream upper-level disturbances originating in the western Eurasian continent (Joung and Hitchman 1982; Wu and Chan 1997; Chen et al 2002). However, this possible influence from the upper troposphere to the mechanism of the cold air outbreaks has not been fully explored and which is addressed here.

Figure 5.5 shows the monthly mean winds (m/s) of JS at 250 hPa derived from the daily winds at 00, 06, 12 and 18 UTC as available from ECMWF, ERA Interim for the month of Jan 2016, Dec 2009 and Jan 2008. The JS stretches from East Asia to the northern Pacific (defined here for the region between Lat 25°N - 45°N and Lon 70°E - 180°E) with a peripheral wind moving eastward. The JS reaches a maximum monthly wind speed of 70-80 m/s at its core for Jan 2016 and Jan 2008 while it was weaker with wind speed maximum of 60-70 m/s for Dec 2009. The inner core of the JS also referred to as 'Jet Streak' extends in longitude from around 130°E up to 180°E and is confined between the latitude from 25°N to 45°N and is associated with a relatively high monthly mean wind of 80 m/s for the Jan 2016 case. The Jet Streak shrunk in longitude extending from 125°E - 155°E for Jan 2008 and did not exist for Dec 2009. In Section 5.4, we investigate the connection of the SMH, AL and JS for signals of strong CS propagating towards East Asia region.

Figure 5.3 Monthly averaged SMH MSLP (a) Jan 2016 (b) Dec 2009 (c) Jan 2008 and (d) Feb 2008 computed from ECMWF, ERA Interim analysis field based on daily maximum values over 00 UTC, 06 UTC, 12 UTC, and 18 UTC. The coastal boundaries of Lake Baikal are represented by black thick line. The domain indicates the defined SMH.

Figure 5.4. Monthly averaged AL MSLP (a) Jan 2016 (b) Dec 2009 (c) Jan 2008 and (d) Feb 2008 computed from ECMWF, ERA Interim analysis field based on daily min values over 00 UTC, 06 UTC, 12 UTC, and 18 UTC. Coastal boundaries are represented by black thick line. The domain indicates the defined AL.

Figure 5.5. Monthly mean wind plots for JS (a) Jan 2016 (b) Dec 2009 (c) Jan 2008 and (d) Feb 2008 computed from ECMWF, ERA Interim based on daily values at 00 UTC, 06 UTC, 12 UTC, and 18 UTC. The coastal boundaries are represented by black thick line.

5.4 Identification of a unique pattern related to the onset of CS

A fundamental objective of synoptic meteorology is to relate the atmospheric circulation to that of the surface environment (Yarnal, 1993) which is a result of both local and large scale forcing. Here we postulate in Section 5.4.1 the outbreak of CS from the SMH in relationship to the intensification of low pressure over AL through the variability of pressure difference between these two systems. The pressure difference is mainly responsible for accelerating a parcel of air from a region of high pressure to a region of lower pressure region, resulting in wind. Thus, a greater pressure difference (PD) over a given horizontal distance, the stronger is the wind. However, for large scale flows, apart from pressure difference there are also other forces acting, such as surface friction force (resulting in formation of Ekman Spiral), Coriolis force generally balancing the pressure difference (as following the Buys-Ballot law) and centrifugal force. Here we postulate that there is also a significant influence from the upper troposphere through the JS that contributes to the cold air outbreaks in East Asia as originating from the synoptic disturbances at the lower troposphere, i.e. from SMH and AL.

5.4.1 Role of SMH and AL systems

Figure 5.6 shows the daily MSLP averaged over the SMH domain during Jan 2016, Dec 2009, Jan & Feb 2008 with intraday variations at 00, 06, 12 and 18 UTC of each day shown. It shows that for Jan 2016, the SMH domain exhibited a strong overall increasing MSLP (MK test with Z statistics of 5.08 and p value of 1.814×10^{-6}) (Figure 5.6a). The MSLP rises steeply from 1033.13 hPa on 18th Jan 06 UTC and reaching the month's maximum value of 1055.29 hPa on 22nd Jan 00 UTC, thus, indicating the presence of massive cold air over the SMH domain. From 22nd Jan 06 UTC, the MSLP starts falling consistently till 26th Jan 12 UTC. The SMH intensity is found to be strongest for the month of Jan 2016 with the maximum attained value of 1055.29 hPa, followed by Jan 2008 (1047.2 hPa), Dec 2009 (1043.18 hPa) and Feb 2008 (1039.64 hPa). For Jan 2016, there was also a sudden and steep increase in the MSLP over four consecutive days from

1033.13 hPa on 18th Jan 06 UTC (Figure 5.6a) before reaching the month's max attained MSLP of 1055.29 hPa on 22nd Jan at 00 UTC. The MSLP then started to weaken from 1054.303 hPa on 22nd Jan at 06 UTC and drastically dropped to 1028.34 hPa on 26th Jan 18 UTC. For Dec 2009, an overall increasing trend in MSLP is observed (trendline using MK test with Z statistics of 1.35 and p value 0.176). Furthermore, a consistent increase in MSLP is observed from 1031.3 hPa on 12th Dec 06 UTC to 1041.15 hPa on 14th Dec 18 UTC, this then dropped for nearly 36 hr before increasing to reach the month's maximum attained MSLP of 1043.187 hPa on 17th Dec at 18 UTC (Figure 5.6b). For Jan 2008, an overall increasing trend in MSLP is observed (MK test with Z statistic 3.03 and p value of 94×10^{-10}). MSLP starts to increase consistently from 1026.77 hPa from 9th Jan 06 UTC before reaching the month's max attained MSLP of 1047.2 hPa on 12th Jan 00 UTC (Figure 5.6c). Feb 2008 indicated an overall weakening of the MSLP (MK test with Z statistic of (-) 8.40 and p value of 9.97×10^{-21}) in contrast to the strengthening trend for Jan 2016, Dec 2009 and Jan 2008 and attained the month's maximum attained MSLP of 1039.64 hPa on 11th Feb at 00 UTC, and with strengthening started from 1031.88 hPa on 9th Feb 06 UTC (Figure 5.6d). Thus, in all these cases the MSLP over SMH started to strengthen consistently within 48 to 108 hr before reaching the month's maximum attained value.

Figure 5.6. MSLP (hPa) over SMH domain for (a) Jan 2016; (b) Dec 2009; (c) Jan 2008; (d) Feb 2008 over the entire month, computed from ECMWF, ERA Interim with intraday values at 00, 06, 12, 18 UTC. Dotted lines in (a) to (d) indicates fitted linear trends.

The AL intensity in Figure 5.7 (which shows the daily MSLP averaged over the AL domain) is found to be strongest for the month of Jan 2016 with MSLP dipping to the lowest value of 996.89 hPa, followed by Dec 2009 (999.36 hPa), Jan 2008 (1002.58 hPa) and Feb 2008 (1000.29 hPa). For Jan 2016 there is an overall increasing MSLP using MK test with Z statistic of 2.32 and p value of 0.363. Furthermore, there is a sudden and steep decrease in the MSLP for six consecutive days from 1011.78 hPa on 20th Jan at 06 UTC before dipping to the month's min MSLP of 996.83 hPa on 26th Jan 06 UTC (Figure 5.7a). The MSLP then starts to strengthen from 997.02 hPa on 22nd Jan 12 UTC and rapidly rises to 1009.45 hPa on 31st Jan 18 UTC. For Dec 2009 there is an overall decreasing trend in MSLP (MK test with Z statistic of (-) 7.16 and p value 1.14×10^{-15}). MSLP dips to a minimum value of 999.36 hPa on 27th Dec 06 UTC and then rapidly rises to 1009.45 hPa on 31st Jan 18 UTC (Figure 5.7b). For Jan 2008, there is an overall increasing trend in MSLP (MK test with Z statistic of 8.60 and p value of 3.83×10^{-20}). MSLP dips to a minimum value 1002.587 hPa on 3rd Jan at 18 UTC and consistently starts strengthening from 1002.62 hPa from 4th Jan 00 UTC and reaches the month's maximum value 1021.68 hPa on 24th Jan 06 UTC from whence it starts dropping drastically to a value 1011.74 hPa on 28th Jan 12 UTC (Figure 5.7c). Whilst, for Feb 2008, there is an overall decreasing trend in MSLP (MK test with Z statistic (-) 2.37 and p value 0.083). MSLP dipped to its minimum value of 1000.29 hPa on 18th Jan 06 UTC and started to increase rapidly till it reached 1015.76 hPa on 24th Feb 06 UTC (Figure 5.7d). Thus, over all these cases for Jan 2016, Dec 2009 and Jan-Feb 2008 the AL MSLP attains the smallest month's max (1011 hPa) for Jan 2016 case on 20th Jan 2016.

Figure 5.7. MSLP (hPa) over AL for (a) Jan 2016; (b) Dec 2009; (c) Jan 2008; (d) Feb 2008, computed from ECMWF, ERA Interim with intraday values at 00, 06, 12, 18 UTC shown. Dotted lines in (a) to (d) indicates fitted linear trends.

Figure 5.8 shows the daily PD between the SMH and AL domains during Jan 2016, Dec 2009, Jan & Feb 2008 with intraday variations at 00, 06, 12 and 18 UTC shown. The PD is found to be strongest for the month of Jan 2016 with the maximum attained value of 49.69 hPa, followed by Dec 2009 (34.45 hPa), Jan 2008 (34.07 hPa) and Feb 2008 (33.09 hPa). For Jan 2016, an overall increasing trend in PD is observed (MK test with Z statistic of 3.64 and p value of 1.32×10^{-5}). Furthermore, there is a sudden and steep increase in the PD from 24.85 hPa on 18th Jan 12 UTC for the three consecutive days before rapidly increasing to reach the month's max attained PD of 49.69 hPa on 22nd Jan 18 UTC (Figure 5.8a). The PD then starts to weaken from 49.43 hPa on 23rd Jan 00 UTC and drastically drops to 31.32 hPa on 26th Jan 12 UTC. For Dec 2009 an overall increasing trend in PD is observed (MK test with Z statistic of 4.55 with p value of 8.35×10^{-7}). A consistent increase in PD is observed from 19.68 hPa on 12th Dec at 18 UTC to 31.91 hPa on 16th Dec 00 UTC. It then dropped consecutively for 12 hr and then increased to reach the month's max attained PD of 34.45 hPa on 17th Dec 18 UTC (Figure 5.8b). For Jan 2008, an overall increasing trend in PD is seen (MK test with Z statistic of 0.52 and p value of 0.163). The PD started to increase consistently from 15.49 hPa from 8th Jan 06 UTC before reaching the month's max attained PD of 34.07 hPa on 12th Jan 00 UTC (Figure 5.8c). Feb 2008 had an overall weakening of the PD (MK test with Z statistic (-) 4.43 and p value of 6.45×10^{-8}) in contrast to the strengthening trend seen for Jan 2016, Dec 2009 and Jan 2008. However, the month's maximum attained PD on 18th Feb at 00 UTC had strengthening from 24.56 hPa on 16th Feb 06 UTC (Figure 5.8d). Thus, in all these cases the PD started to strengthen consistently within 48 to 108 hr before reaching the month's maximum attained value.

Figure 5.8. PD (hPa) for (a) Jan 2016; (b) Dec 2009; (c) Jan 2008; (d) Feb 2008 between the SMH domain and the AL domain over the entire month, computed from ECMWF, ERA Interim at 00, 06, 12, 18 UTC. Dotted lines in (a) to (d) indicates fitted linear trends.

During Jan 2016, the PD started to strengthen from 18th Jan 12 UTC and attained its maximum on 22nd Jan 18 UTC (Figure 5.8a). This is due to the intensification of AL, first driving winds from the Arctic Circle through Eastern Russia and towards the AL system (Figure 5.9a). Simultaneously, due the strengthening of the SMH (Figure 5.7a), the PD rapidly intensifies to the month's maximum value shown in Figure 5.8a on 22nd Jan 2016 18 UTC. This PD drove a wide continuous band of winds having speed of 24-27 m/s (Figure 5.9b, brownish color) moving consistently over the Arctic Circle through the North Pacific Ocean, and entering the AL region at a speed of 18-21 m/s (Figure 5.9b, greenish color). However, it is also seen in Figure 5.9a, there is another cold wind trajectory on 18th Jan 12 UTC from the mainland China into the Yellow Sea. This was due to the formation of a localized intense and short lived (24-48 hr) LPS with a central pressure of 992 hPa situated at the eastern coast of Japan, which later weakened as discussed in Section 4.4. Similarly, For Dec 2009, the PD started to strengthen from 12th Dec 18 UTC and reached the maximum value on 17th Dec 18 UTC (Figure 5.8b). For Jan 2008, the PD strengthened from 8th Jan 06 UTC and reached maximum value on 12th Jan 00 UTC (Figure 5.8c) while for Feb 2008, it strengthened from 16th Feb 00 UTC and reached maximum value on 18th Feb 06 UTC (Figure 5.8d). In all these cases, the PD started to strengthen due to the intensification of the AL, driving the winds from Arctic Circle through eastern Russia towards the AL system (Figure 5.9 c, e, and g). Due to the strengthening of SMH, the PD rapidly intensified to reach the maximum value developed between SMH and AL systems. This drove the winds from the Arctic Circle through the north Pacific Ocean and entered the region dominated by AL with a wind speed of 10-15 m/s in Dec 2009 (Figure 5.9d), 10-15 m/s in Jan 2008 (Figure 5.9f) and 15-18 m/s in Feb 2008 (Figure 5.9h).

It is also noted that in all the three cases, the cold wind trajectory is seen to be initially moving from mainland China into the Yellow Sea (Figure 5.9 a, c, e and g) due to the formation of a localized, intense and short lived (24-48 hr) intense LPS situated at the eastern coast of Japan. With the consistent weakening of the PD after the monthly maximum, e.g. from 23rd to 26th Jan 2016 (Figure 5.8a); from 17th to 21st Dec 2009

(Figure 5.8b); 12th to 17th Jan 2008 (Figure 5.8c); 18th to 21st Feb 2008 (Figure 5.8d), the continuous band of wind that originated over the Arctic Circle became distorted and discontinuous over the region of the Russian mainland to Mongolia. They then entered the AL region through the Northern Pacific Ocean with a reduced wind speed during Jan 2016 (Figure 5.10a), Dec 2009 (Figure 5.10b), Jan 2008 (Figure 5.10c) and Feb 2008 (Figure 5.10d). The above analysis shows that the development of an intense PD between SMH and AL significantly dominated the progression of a continuous and intense band of winds from the Arctic Circle, reaching far into the lower latitudes.

Figure 5.9. ECMWF, ERA Interim 925 hPa winds (m/s) for (a) 12 UTC 18th Jan 2016; (b) 18 UTC 22nd Jan 2016; (c) 18 UTC 12th Dec 2009; (d) 18 UTC 17th Dec 2009 (e) 06 UTC 8th Jan 2008; (f) 00 UTC 12th Jan 2008; (g) 06 UTC 16th Feb 2008; (h) 00 UTC 18th Feb 2008. The red box indicates the combined SMH and AL domains.

Figure 5.10. ECMWF, ERA Interim 925 hPa winds (m/s) for (a) 00 UTC 26th Jan 2016; (b) 00 UTC 21st Dec 2009; (c) 00 UTC 17th Jan 2008; (d) 00 UTC 21st Feb 2008. The red box indicates the combined SMH and AL domain.

5.4.2 Role of JS and Jet Streak

In Section 5.4.1 we infer that at the surface, CS are characterized by outbreak of cold air masses with strengthening of SMH and its progression towards the lower latitudes through intensification of MSLP at AL. During the EAWM, the strong westerly JS, specifically the PJ dominates the upper troposphere circulation system in over East Asia. Previous literatures indicated that the CS during the EAWM are often preceded by upstream upper-level disturbances originating in the western Eurasian continent (Joung and Hitchman 1982; Wu and Chan 1997; Chen et al 2002). However, this possible influence from the upper troposphere to the mechanism of the cold air outbreaks has not been fully explored and which is addressed here.

A further analysis of the JS velocity averaged over its defined region of Lat 25°N - 45°N and Lon 115°E - 180°E over East Asia at daily level shows that the JS had an overall increasing trend during Jan 2016 and Dec 2009, while it had a decreasing trend for Jan 2008 (MK test with Z statistic of 3.49, 3.99, (-) 2.71 and p values of 6.683×10^{-6} , 6.16×10^{-7} , 0.043 respectively) (Figure 5.11). Furthermore, the intensity of the JS started increasing rapidly and consistently for 8 days from 51.97 m/s on 19th Jan 00 UTC to its maximum attained of 67.66 m/s on 27th Jan 18 UTC (Figure 5.11a, red box). For Dec 2009, the JS intensity started increasing continuously for 6 days from 45.83 m/s on 13th Dec 18 UTC and reached its maximum value of 59.88 m/s on 19th Dec 18 UTC (Figure 5.11c, red box). For Jan 2008, the JS increased rapidly for 5 consecutive days from 53.07 m/s on 4th Jan 18 UTC to 60.06 m/s on 9th Jan 06 UTC (Figure 5.11e, red box). Notably, for Jan 2016, the period when the JS was strengthening coincided with a steep consistent drop in MSLP over AL with the MSLP value dipping from 1011.81 hPa on 20th Jan 06 UTC to the month's lowest attained value of 996.83 hPa (Figure 5.11b, red box). The same behavior is observed for Dec 2009. During the intensification of JS, there was a continuous drop in the AL MSLP from 1013.612 hPa on 13th Dec 18 UTC to 1008.34 hPa on 18th Dec 00 UTC (Figure 5.11d, red box). Similarly, for Jan 2008, during the strengthening of JS, a sudden and drastic drop of 12 hPa in MSLP over AL was observed

from 1014.45 hPa 1st Jan at 00 UTC 1002.5 hPa to 3rd Jan 18 UTC after which it started rising gradually till 9th Jan at 18th UTC but remained below 1012 hPa (Figure 5.11f, red box). Thus, in the above cases the strengthening of the JS was accompanied by the intensification of AL. The AL started to intensify, nearly a day after the strengthening of Jet Streak and the intensification weakened a day before the weakening of the Jet Streak.

Vertical cross sections of the JS are shown in Figure 5.12. The cross section along the Jet Streak extended from surface to the upper air at 250 hPa along the longitudes. The vertical wind speed reached its maximum speed of $27 \text{ Pa/s} \times 10^{-1}$ for Jan 2016 (Figure 5.12a, red box) whilst for Jan 2008 it was limited to $15 \text{ Pa/s} \times 10^{-1}$ but had a larger spatial extent of intense convection influencing the area between 125°E - 140°E (Figure 5.12c, red box). The daily level analysis of the JS showing the origination of the ‘Jet Streak’ embedded within the JS during its strengthening phase is shown in Figure 5.13a for 19th Jan 2016 00 UTC, 13th Dec 2009 18 UTC (pinkish color, Figure 5.13b), and 4th Jan 2008 18 UTC (pinkish color, Figure 5.13c). As the JS progressed, eastward, it reached its maximum spatial strength and extended longitudinally over 40° i.e. from 125°E to 165°E for Jan 2016. It then moves over South Korea to the Japan region on 27th Jan 18 UTC (Figure 5.14a). Similarly, the Jet Streak strengthens as it progresses eastward and extends longitudinally till it reaches its maximum strength on 19th Dec 2009 18 UTC (Figure 5.14b) and 9th Jan 2008 06 UTC (Figure 5.14c). The vertical cross section of the Jet Streak for 27th Jan 2016 along its center at 37°N (Figure 5.15a) shows intense ascending winds (red color) with speed $6 \text{ Pa/s} \times 10^{-1}$ between 120°E to 140°E and intense descending winds (purple color) between 140°E to 180°E with speed $(-) 4 \text{ Pa/s} \times 10^{-1}$. Similarly, for 19th Dec 2009 and Jan 2008, the vertical cross section along the Jet Streak shows intense converging winds ascending with a maximum speed of $6 \text{ Pa/s} \times 10^{-1}$ (Figure 5.15b) and $8 \text{ Pa/s} \times 10^{-1}$ (Figure 5.15c) respectively and diverging winds with a maximum descending speed of $5 \text{ Pa/s} \times 10^{-1}$ (Figure 5.15b) and $6 \text{ Pa/s} \times 10^{-1}$ (Figure 5.15c) respectively.

Air parcels moving at different speeds within the exit and entrance regions of Jet Streak become out of balance due to the existing temperature gradient prevailing in these

regions. This balance is restored by the atmosphere through vertical motion which is attained through ageostrophic winds (Uccellini and Kocin, 1987). The eastward flowing air as it enters the Jet Streak, it accelerates vertically upwards towards low pressure across the height contours and this creates convergence associated with rising air parcels. This is maximized on the longitudes below 155°E . Within the JS entrance and exit regions, the ageostrophic wind dominates the amount of divergence/convergence (and subsequent vertical motion) due to jet streak dynamic (Bjerknes (1951). Whereas, the air leaving the Jet Streak creates divergence associated with sinking air parcels and is maximized on the longitudes greater than 155°E . The persistent pattern of intensification of AL with the strengthening of Jet Streak during the month of Jan 2016, Dec 2009 and Jan 2008 (Figure 5.11) and convergence at the surface and upper level features (Figure 5.12 and Figure 5.15) suggests that vertical transverse circulations associated with the upper level Jet Streak at 250 hPa may be a link that drives the progression of CS from SMH towards the lower latitudes, significantly effecting the regions within and underneath the converging zone of Jet Streak, and leading to the formation of intense low pressure systems in that region. This hypothesis is next tested by examining the JS flow during the CSE.

Figure 5.11. Six hourly variation of JS (m/s) in (a), (c), (e) and AL (hPa) in (b), (d), (e) from ECMWF, ERA Interim from (a) & (b) 1st Jan-31st Jan 2016; (c) & (d) 1st Dec-31st Dec 2009; and (e) & (f) 1st Jan-31st Jan 2008. Dotted lines in (a), (c) & (e) indicates fitted linear trends.

Figure 5.12. Vertical Winds between 1000 hPa and 250 hPa, computed from ECMWF, ERA Interim for the month of (a) Jan 2016 at Lat 32°N ; (b) Dec 2009 at Lat 31°N ; and (c) Jan 2008 at Lat 32°N . The red boxes represent the intense vertical winds ((1/10) *Pa/s) along the cross section. Positive and negative values indicate rising and sinking air respectively.

Figure 5.13. Jet Streak (solid box) within the JS (m/s) at 250 hPa, computed from ECMWF, ERA Interim valid for (a) 00 UTC 19th Jan 2016; (b) 18 UTC 13th Dec 2009; and (c) 18 UTC 4th Jan 2008.

Figure 5.14. Jet Streak (solid box) within the JS (m/s) at 250 hPa, computed from ECMWF, ERA Interim valid for (a) 27th Jan 2016 18 UTC; (b) 19th Dec 2009 18 UTC; and (c) 9th Jan 2008 06 UTC.

Figure 5.15. Vertical Winds between 1000 hPa and 250 hPa, computed from ECMWF, ERA Interim valid for (a) 18 UTC 27th Jan 2016 at Lat 37° N; (b) 18 UTC 19th Dec 2009 at Lat 30° N; and (c) 06 UTC 9th Jan 2008 at Lat 33° N. The red boxes represent the intense vertical winds ((1/10) *Pa/s) along the cross section.

Figure 5.16 shows the flow of the JS during the outbreak of the strong CS on Jan 2016, Dec 2009 and Jan-Feb 2008. The JS at 250 hPa at locations closer to HK on 24th Jan 2016 00 UTC has a speed of around 90 m/s (Figure 5.16a, deep brown), indicating a rapid progression of large cold air from the SMH toward HK from the mainland China. The JS affects after crossing over HK region become less intense with a reduced speed of 70 m/s (Figure, 5.16b light brown). For Dec 2009, a weaker JS with a speed of around 70 m/s (Figure, 5.16c, yellowish color) is observed when it passed over HK region. As with Jan 2016, the continuous jet streak (yellowish color) over HK indicated a rapid progression of cold air from the SMH through mainland China towards HK. The Jet Streak embedded within the JS strengthened with a speed of around 80 m/s (Figure 5.16d, dark brown) as it moved eastward. For the CSE Jan 2008, the Jet streak within the longitude 130-180 °E and latitude 25-45 °N started to strengthen (Figure 5.16e, yellowish color) from 26th Jan 2008 at 00 UTC and continues to strengthen till 15th Feb 2008 18th UCT (Figure 5.16f, pinkish color). In the above three cases, the pattern of the JS flow is observed to be zonal (nearly parallel to the latitudes) passing over East Asia and had a long continuous wave extending from 70 °E to beyond 180 °E with a further embedded Jet Streak. The continuous and stronger JS indicated a large temperature difference with a rapid change in temperature taking place over the defined domain of the Jet Streak within the longitude 130-180 °E and latitude 25-45 °N. Thus, the strengthening of the Jet Streak as embedded within the JS is coincident with a rapid progression of CS from the SMH and was also associated with intense convection in the atmospheric column underneath the Jet Streak.

For the CSE during Jan-Feb 2008, a very distinct pattern of zonal JS is observed where the embedded Jet Streak starts strengthening from 26th Jan 00 UTC that continues till 15th Feb 12 UTC (Figure 5.16e & f). An elongated continuous stretch of JS is observed extending from 70 °E to 180°E with an embedded Jet Streak located within longitude of 130-180 °E. This originated on 26th Jan 2008 00 UTC and remained till 15th Feb 12 UTC. This JS later started to split from 15th Feb at 18 UTC (Figure 5.16g) and further weakened as it moved eastward. Over this duration, HKO recorded the longest cold spells since 1968 that started from 24th Jan and ended on 16th Feb 2008. This further indicated a rapid

progression of cold air from the SMH at an upper level underneath the Jet Streak which is within the domain defined for the formation of LPS at surface (Section 4.4). Thus, in addition to the strength of Jet Streak, the pattern of JS itself plays a significant role in the onset of CS towards lower latitudes.

ECMWF 250 hPa Winds (m/s) at 00 UTC on 24th
Jan 2016

ECMWF 250 hPa Winds (m/s) at 12 UTC on 24th
Jan 2016

ECMWF 250 hPa Winds (m/s) at 12 UTC on
15th Dec 2009

ECMWF 250 hPa Winds (m/s) at 00 UTC on
16th Dec 2009

ECMWF 250 hPa Winds (m/s) at 00 UTC on 26th
Jan 2008

ECMWF 250 hPa Winds (m/s) at 12 UTC on
15th Feb 2008

Figure 5.16. ECMWF, ERA Interim plot of JS over East Asia towards Northern Pacific at 250 hPa valid on (a) 24th Jan 2016 00 UTC; (b) 24th Jan 2016 12 UTC; (c) 15th Dec 2009 12 UTC; (d) 16th Dec 2009 00 UTC; (e) 26th Jan 2008 00 UTC; (f) 15th Feb 2008 12 UTC; and (g) 15th Feb 2008 18 UTC.

5.4.3 Formation of LPS near Coast of Japan

In Section 5.4.2, we demonstrated that the persistent pattern of intensification of AL with the strengthening of Jet Streak as embedded within the JS, specifically the PJ played a significant role in the onset of CS towards lower latitudes. We also discussed on the possibility of the formation of LPS underneath the zone of Jet Streak and dominance of the LPS in determining the CS trajectories (Section 4.4). In this section, we further investigate the mechanism that leads to LPS and to the authors knowledge this is reported for the first time on the role of JS in inhibiting the formation and development of these intense short lived LPSs.

Figures 5.17, 5.18, 5.19 and 5.20 shows the condition corresponding to the strongest Jet Streak during the month of Jan 2016, Dec 2009 and Jan-Feb 2008. All have wind speeds above 100 m/s (pinkish color) at 250 hPa and are located between latitude 25-40 °N and

longitude 120-180° E (sub-figures (a) of Figures 5.17, 5.18, 5.19 and 5.20). The vertical cross sectional structure of the Jet Streak and the formation of LPS off the eastern coast of Japan are shown in sub-figures (b) and (c) of Figures 5.17, 5.19, 5.20 and sub-figure (b), (c) and (d) of Figure 5.18. For Jan 2016, the jet streak extended over the eastern coast of Japan towards the Pacific Ocean within 33-39°N and 145-165°E (Figure 5.17a). The vertical cross section along the center of the Jet Streak at 36°N (Figure 5.17b) shows surface level ascending air with a wind speed of $6 \text{ Pa/s} \times 10^{-1}$ and an intense divergence at the exit of the Jet Streak from the upper level with a speed of above $18 \text{ Pa/s} \times 10^{-1}$. Further, the region of diverging winds in the upper level is approximately three times larger than the convergent indicating instability of air masses in the vertical column due to removal of more air from the column that furthered lowered the surface pressure and led to the formation of an intense LPS with central pressure of 996 hPa (Figure 5.17c). For Dec 2009, we see an amplified Jet Streak towards north with spatial extent extending over eastern coast of Japan and into the Pacific Ocean (140°E-175° W) (Figure 5.18a). The vertical cross section along the entrance and exit of the Jet Streak (Figure 5.18a) shows higher descending air with speeds of $10 \text{ Pa/s} \times 10^{-1}$ and $16 \text{ Pa/s} \times 10^{-1}$ respectively (Figure 5.18b & 5.18c), leading to the formation of two intense LPSs with central pressure of 990 hPa and 960 hPa (Figure 5.18d). For Jan-Feb 2008 a weaker jet streak is seen as evidenced through its limited spatial longitudinal extent spanning over 15° (Figure 5.19a and Figure 5.20a). Furthermore, we see that the convergence and divergence of the air masses here was mostly governed by the next adjacent layer of the air masses spanning over 15° in longitude with the zonal wind speed of 90 m/s, embedding the Jet Streak (Figure 5.19a and Figure 5.20a). For Jan 2008, air is seen rising at the surface level with $2 \text{ Pa/s} \times 10^{-1}$ and descending at $8 \text{ Pa/s} \times 10^{-1}$, as evidenced from the vertical cross section of the Jet Streak (Figure 5.19b). However, under the influence of this next adjacent layer of the air masses spanning over 15° in longitude the intensity at the ascending air at surface level increased to $12 \text{ Pa/s} \times 10^{-1}$ at the surface while at the exit the descending air at the upper level also increased to 1.0 Pa/s impacting a larger area at the surface level within 165° E -175° E (Figure 5.19b). This led to an intensified LPS with 998 hPa as the central pressure (5.19c). Similarly, for Feb 2008 below the exit of the Jet Streak, ascending air

with $5 \text{ Pa/s} \times 10^{-1}$ at the surface and descending air of $20 \text{ Pa/s} \times 10^{-1}$ in the upper air (Figure 5.20b) led to the formation of LPS with 980 hPa central pressure at the surface (Figure 5.20c). This analysis further confirms the observations made in Section 5.4.2, that the pattern of JS and strength of Jet Streak is a key factor in determining the intensity and location of LPS which in turn governs the trajectory of CS at the surface (Section 4.4). In addition, to the above cases, we also examined the possible formation of such LPS under the similar influence from JS on various other days during Jan 2016, Dec 2009 and Jan-Feb 2008. For instance, the presence of Jet Streak (between 70-90 m/s) led to the formation of LPS during Jan 2016 on 1st, 5th, 7th, 11th, 12th, 13th, 15th, 19th, 20th, 25th and 31st (Appendix, Figure D.1); during Dec 2009 on 2nd, 4th, 7th, 13th, 14th, 16th, 17th, 18th, 19th, 20th, 21st, 22nd, 27th, 28th, 29th, 30th and 31st (Appendix, Figure D.2); during Jan 2008 on 6th, 7th, 9th, 10th and 30th and during Feb 2008 on 3rd, 4th, 7th, 8th, 10th, 11th and 13th (Appendix, Figure D.3). We therefore, categorize the JS into two sub categories, one with fully developed Jet Streak i.e. spanning over 15° in longitude (i.e. Jan 2016 and Dec 2009 case) and the other with a partially developed jet streak i.e. spanning less than 15° along the longitude (i.e. Jan & Feb 2008 case).

Figure 5.17. (a) JS at 250 hPa; (b) Vertical winds at Lat 36°N between 1000 hPa and 250 hPa; (c) MSLP plot from ECMWF, ERA Interim for 00 UTC on 24th Jan 2016. Dotted white line in (a) represents line joining entrance and exit of the Jet Streak at 36°N. The red arrows in (b) indicates the ascending / descending air masses.

Figure 5.18. (a) JS at 250 hPa; (b) Vertical winds at Lat 34°N between 1000 hPa and 250 hPa ; (c) Vertical winds at Lat 35°N between 1000 hPa and 250 hPa ; (d) MSLP plot computed from ECMWF, ERA Interim valid for 00 UTC on 21st Dec 2009. Dotted white line in (a) represents the line at the entrance and at the exit of the Jet Streak at 34°N and 35°N respectively. The red arrows in (b) and (c) indicates ascending / descending air masses.

Figure 5.19. (a) JS at 250 hPa; (b) Vertical winds at Lat 36°N between 1000 hPa and 250 hPa; (c) MSLP plot computed from ECMWF, ERA Interim valid for 06 UTC on 17th Jan 2008. Dotted white line in (a) represents the line joining the entrance and the exit of the Jet Streak at 36. The red arrows in (b) indicate ascending /descending air masses.

Figure 5.20. (a) JS at 250 hPa; (b) Vertical winds at Lat 36°N between 1000 hPa and 250 hPa; (c) MSLP plot computed from ECMWF, ERA Interim valid for 00UTC on 13th Feb 2008. Dotted white line in (a) represents the line joining the entrance and the exit of the Jet Streak at 36. The red arrow in (b) indicate ascending / descending air masses.

5.5 Discussion & Summary

Our analysis in Section 5.4 confirms that the intensity and progression of CSEs for Jan-Feb 2008, Dec 2009 and Jan 2016 in East Asia are significantly influenced by the interactions of the three large atmospheric systems i.e. SMH, AL and JS. Figure 5.21 shows the monthly averaged SMH and AL over 38 years for the months of Dec, Jan and Feb, derived from the daily analysis field of ECMWF, ERA Interim over 1979-2017. The strongest SMH at MSLP of 1040.35 hPa occurred during Jan 2011 whereas the most intensified AL with MSLP of 1001.51 hPa occurred during Jan 1981 (Figure 5.21a). The strongest PD between the SMH and AL systems of 33.21 hPa occurred during Jan 2016 followed by 32.44 hPa in Jan 1981 (Figure 5.21b). Similarly, for the month of Dec the strongest SMH is 1037 hPa and the most intensified AL is 1005 hPa occurring during years 1986 and 2005 (Figure 5.21c), respectively and with the strongest PD of 31.38 hPa occurring in 2005 (Figure 5.21d). The month of February remains less amplified compared to the months of Jan and Dec. The SMH reaches the maximum strength of 1034 hPa in 1986 and AL intensifies to 1003 hPa in 1983 (Figure 5.21e) with the strongest PD of 26.26 hPa in 1980 (Figure 5.21f). It is observed that there is an overall increasing trend in the strength of SMH system for Jan in the past 38 years whereas there is no significant trend observed for the AL system over the same period. It is further noted that strengthening trend of SMH is more evident for the later period 1996 to 2017 whereas there is a weakening trend from 1979 to 1995. A further examination of the monthly averaged MSLP over 38 years for the months of Dec, Jan and Feb shows that the highest SMH monthly average MSLP value of 1031.09 hPa is attained during the month of Jan, followed by Dec and Feb (Table 5.1). During these years the most intensified AL reached an average MSLP value of 1009.87 hPa during the month of Jan, followed by Dec and Feb (Table 5.1). The strongest PD of 21.22 hPa between the SMH and AL systems also occurred for the month of Jan followed by Dec and Feb (Table 5.1). Furthermore, an increasing trend for PD is observed for Jan while decreasing trends for Dec and Feb (MK test with Z statistic of 0.24, (-)0.03, (-)0.38 and p value of 0.545, 0.938, 0.914, respectively) (Figure 5.21 b, d, & e). This shows that Jan is the most critical month for the

intense progression of the cold air masses from SMH. The comparison of the daily meteorological variables with the monthly values during CSEs for Jan-Feb 2008, Dec 2009 and Jan 2016 are presented next. Tables E.1, E.2, E.3 & E.4 in Appendix E list the daily and normal (monthly mean over 1979-2016) PD and JS values. For CSE 2016, the daily mean PD for Jan 2016 was 33.1 hPa which was higher than its climatological normal of 21.1 hPa. During the onset of the CS, the PD and JS continually strengthened from 21st to 24th Jan. For CSE 2008, the daily mean PD for Jan 2008 was 3.5 hPa higher than its climatological normal. During this month the PD and JS continually strengthened twice from 1st to 4th and from 18th to 20th Jan. This was also observed during Feb 2008 over 9th-11th when PD and JS continually started to strengthen. During this month the daily mean PD was 4 hPa higher than its climatological normal of 16.6 hPa. For CSE 2009 the daily mean PD for Dec 2009 was 2 hPa which was higher than its climatological normal of 18.9 hPa. During the onset of the CS, the PD and JS continually strengthened from 13th to 16th Dec.

Figure 5.21. MSLP (hPa) over SMH and AL ((a), (c), (e)) and PD (hPa) between SMH and AL ((b), (d), (f)) for Jan, Dec and Feb computed from ECMWF ERA Interim (38 years of daily data over 1979-2017). Dotted lines in (a) to (f) indicates fitted linear trends.

Table 5.1 Monthly average SMH, AL and their PD computed over period 1979-2017

Months	Monthly averaged SMH (hPa) from 1979-2017	Monthly averaged AL (hPa) from 1979-2017	Monthly average PD between SMH and AL (hPa) from 1979-2017
Dec	1029.85	1010.86	18.98
Jan	1031.09	1009.87	21.22
Feb	1028.32	1011.64	16.67

Figure 5.22 shows the PD for Dec-Jan-Feb (blue, red and green respectively) from 1979 to 2016. Jan 2016 had the strongest PD of 33 hPa in the last 38 years followed by Jan 1981. For Dec, highest PD of 31 hPa was observed in 2005 and for Feb highest PD of 26 hPa was observed in 1980. This further confirms that Jan is the most crucial month for the outbreak of CSEs followed by Dec and Feb. Hence, higher values of PD attained during a particular month is an indicative of the possible outbreak of CSE. As discussed in Section 5.2 and demonstrated in Section 5.4.1, the PD between SMH and AL significantly dominated the progression of a continuous and intense band of winds from the Arctic Circle, reaching far into the lower latitudes. Hence, we postulate that the monthly averaged PD values specifically 21.22 hPa for Jan, 18.98 hPa for Dec and 16.67 for Feb (Table 5.1) could be used as a criteria for the progression of cold air mass from SMH towards the lower latitude. Based on this proposed criteria, possible CSEs years for months of Dec, Jan and Feb are listed in Table 5.2. In total this gives 55 months over the 39 years during which CSEs could form.

Figure 5.22. Monthly averaged PD (hPa) computed from ECMWF, ERA Interim for the period 1979-2016. In the figure Dec is indicated by blue bar, Jan by red and Feb by green.

Table 5.2 Possible CSEs years for Dec, Jan and Feb based on criteria using the monthly averaged PD from 1979-2017.

Dec		Jan		Feb	
Year	PD (hPa)	Year	PD (hPa)	Year	PD (hPa)
2004	19.21	2005	21.41	1981	16.89
2009	19.38	1984	21.76	1987	17.00
1984	19.52	1987	21.86	1993	17.80
1982	20.12	1983	23.05	1997	17.95
2012	20.32	1986	23.25	2010	18.50
1981	20.75	2003	23.40	1984	19.66
1991	21.59	2012	23.45	1998	20.46
1986	21.66	2001	23.61	2008	20.55
1993	21.75	1985	23.89	1991	20.96
2006	21.87	1992	24.18	1999	21.01
2011	22.56	2008	24.52	2005	21.57
2014	22.95	1995	24.98	2012	21.90
1994	23.37	1998	25.53	2000	22.61
1985	23.66	2010	26.25	1996	23.10
2003	24.67	2011	30.67	1983	24.15
1995	24.69	1981	32.45	1986	24.41
2001	26.58	2016	33.22	2016	24.44
2002	26.99			1988	25.34
2005	31.38			1980	26.27

The climatology of the JS, specifically the monthly averaged wind speed over 38 years from 1979-2017 for the months of Dec, Jan and Feb are plotted in Figure 5.23. For the month of Jan, the Jet Streak with wind speed above 70 m/s was embedded within the JS that extended from 125 °E to 165 °E (pinkish color, Figure 5.23a). Whilst for Feb the Jet Streak with wind speed above 70 m/s was partially developed (pinkish color, Figure 5.24a) and embedded under the adjacent layer with wind speed between 60-70 m/s (brown color, Figure 5.24a). For Dec, the Jet Streak was weaker than both Jan and Feb and is confined to 60-70 m/s (brown color, Figure 5.25a). As discussed earlier in Sections 5.3, 5.4.2 and 5.4.3 as the air masses enters the JS it accelerates and creates convergence which we termed here as ‘entrance’. These air masses decelerate after achieving its maximum speed at the core of the JS, creating divergence which we termed here as ‘exit’. This is confirmed by the 38 years of data that has the vertical winds along the cross section of the Jet Streak joining the entrance and the exit. The entrance at 32 °N shows intense convection at the surface with vertical wind speed of above $2 \text{ Pa/s} \times 10^{-1}$ extending up to the entrance of the Jet Streak. The exit has a weaker diverging wind at $0.5 \text{ Pa/s} \times 10^{-1}$ extending from upper air to the surface (Figure 5.23b, 5.24b and 5.25b). The vertical cross section along AB show two entrance regions around the Jet Streak as the left entrance and right entrance regions indicating the warm air rising in the right entrance region and cold air sinking in the left entrance region (Figure 5.23c, 5.24c and 5.25c). Similarly, the cross section along CD shows two exit regions around the Jet Streak indicating the cold air sinking in the right exit region and the warm air rising in the left exit region (Figure 5.23d, 5.24d and 5.25d). Since convergence adds mass to the air column, the surface pressure generally increased underneath the left entrance of the Jet Streak. Similarly, the surface pressure decreased underneath the right entrance of the Jet Streak because of divergence. Thus, as discussed in Section 5.3 and demonstrated in Section 5.4.2 & 5.4.3, we see a similar signature of JS for Dec, Jan and Feb over 38 years, where the divergence of air masses either at the right entrance or the left exit of the Jet Streak induced low level convergence, acceleration of lower level winds and enhanced rising motion towards the

upper level. This lowered the pressure at the surface causing the formation and intensification of LPS. Thus, this climatological analysis further confirms that the pattern of JS and the strength of embedded Jet Streak determines the formation of LPS that governs the trajectory of CS from SMH at the surface (Section 4.4).

Figure 5.23. (a) Jet Streak embedded within the JS at 250 hPa; (b) Vertical Winds between 1000-250 hPa along the center of the Jet Streak at 32° N lat; and (c) Vertical winds along line AB between 1000-250 hPa at the 'entrance' of Jet Streak; and (d) Vertical winds along line CD between 1000-250 hPa at the 'exit' of the Jet Streak valid for the month of Jan as averaged over 1979-2017, computed from ECMWF, ERA interim.

Figure 5.24 (a) Jet Streak embedded within the JS at 250 hPa; (b) Vertical Winds between 1000-250 hPa along the center of the Jet Streak at 32° N lat; and (c) Vertical winds along line AB between 1000-250 hPa at the 'entrance' of Jet Streak; and (d) Vertical winds along line CD between 1000-250 hPa at the 'exit' of the Jet Streak valid for the month of Feb as averaged over 1979-2017, computed from ECMWF, ERA interim.

Figure 5.25. (a) Jet Streak embedded within the JS at 250 hPa; (b) Vertical Winds between 1000-250 hPa along the center of the Jet Streak at 32° N lat; and (c) Vertical winds along line AB between 1000-250 hPa at the 'entrance' of Jet Streak; and (d) Vertical winds along line CD between 1000-250 hPa at the 'exit' of the Jet Streak valid for the month of Dec as averaged over 1979-2017, computed from ECMWF, ERA interim

Based on our analysis in Sections 4.4 and Sections 5.4.1 to 5.4.3 the combined effects from the upper troposphere through the embedded Jet Streak of JS and the synoptic disturbances at lower troposphere through strengthened SMH and intensified AL induces southward surface cold air outbreak from the eastern SMH leading to strong CSEs in East Asia. This combined impact may further lead to a more severe impactful local weather situation during NEM, where the extreme weather may reach new levels in the regions such as coastal regions of China, regions of North Korea, South Korea and Japan and northern regions of Taiwan. Such regions under significant influence from JS, SMH and AL for Dec, Jan and Feb are shown spatially within the boxes in Figure 5.26. The red box in the figure represents the regions under the influence of intense atmospheric convection at the surface due to the interaction of cold and warm air masses at the lower level and above it, the progression of the JS (Figure 5.26a, b & c). Our analysis in Sections 5.4.1 to 5.4.3 showed that Jan is more crucial for the outbreak of cold air masses followed by Dec and Feb. Thus, convective activity in Jan is represented within two boxes i.e. the red box and violet boxes at the surface (Figure 5.26b) in contrast to Dec and Feb with only red boxes (Figure 5.26a & d). The region within the red boxes represents the regions that are under the influence of intense atmospheric convection at the surface due the progression of a partially developed JS. The violet box for Jan represents the regions under the influence of a fully developed JS (Figure 5.26c).

Figure 5.26. Climatology of MSLP for SMH and AL systems for (a) December; (b) January; and (c) February for the period 1979-2017 as computed from ECMWF, ERA Interim. The red box in (a), (b) and (c) indicates region of active convection due to the progression of JS while the violet box in (b) indicates severe convection due to the progression of JS embedding fully developed Jet Streak during the month of January.

Based on the detailed analysis performed in Sections 4.3, 4.4, & 5.4, we now provide a generalized description of the key meteorological mechanisms and drivers, and the process that causes CSE in East Asia as summarized in the schematic diagram shown in Figure 5.27. Around the start of a CSE, a dominant circulation system is formed as characterized by a strong SMH exceeding its climatological mean in Dec, Jan and Feb respectively. This drives the air masses from Arctic Circle towards central to South China and reaching far into SCS, providing these regions with copious amounts of cold dry air. Another dominant circulation system, the AL centered near Aleutian Islands is characterized by most intense (lowest pressure) below its climatological mean in Dec, Jan and Feb respectively. With intensification of this AL, the cold air masses also start streaming into the same region via Mongolia, East Asia and South China. The PD originating from the strengthened SMH and the intensified AL results in sudden surface temperature drop, leading to strong CSE as observed in Jan 2016 case. Over, East Asia the upper tropospheric flow is characterized by the strengthening of the Jet Streak with JS. The east ward progression of the strengthen Jet Streak leads to intense divergence of air masses at upper air and leads to the formation of intense short lived LPS between coast of Japan and western Pacific (indicated as active convection within red box for Dec & Feb and as severe active convection within violet box). These LPS further intensifies by picking up heat and moisture from the Pacific and leads to two major CS tracks. Associated with the Jet Streak, we found a continuous long JS stretching all the way from Southwest China to Northern Pacific, indicating a large upper air temperature difference. The zonal pattern associated with JS pattern remained stagnant for days leading to colder than average weather over south central China as was evidenced in Jan-Feb 2008 case. Furthermore, the weather situation drastically worsened with the shift in longitudinal shift of the AL system westward from its climatological mean. This westward shift of the AL system allowed cold air from SHM to stream into China, mainly from the west rather than the usual northerly route. These together can lead to severe CSE such as reported in Jan-Feb 2008. Our proposed mechanism thus provides a new theoretical framework for explaining CSEs originating due to prolonged enhanced PD in East Asia. Here we propose the severity of CSE into two categories distinguished by the intensities of PD

using CSE scale. The scale is a categorization scale ranging from ‘Strong’ to ‘Extreme’ for CSEs based on the monthly averaged intensity of PD and strength of Jet Streak derived over a period of 38 years (1979-2016) (Table 5.3). This scale provides an estimate of the potential of CSE and its progression from Siberia. To be classified as strong CSE, a CSE must have a PD of at least 19 hPa. The higher classification, extreme CSE, consists of CSE with PD exceeding at least 29 hPa. Furthermore, the scale also provides a guidance on the progression of CSE. A value of Jet Streak below 67 m/s leads to CSE propagation towards SCS while a high value between 68-77 m/s leads to the bifurcation of CSE track towards SCS and Japan. The highest value of 78 m/s and above majorly dominate the CSE track towards Japan.

Figure 5.27. The progression of continental polar air masses from Arctic towards Siberia as evidenced through strengthening of MSLP. The cold air masses (brown color) as it reaches Mongolia (pinkish color) bifurcates into two major tracks (indicated by black arrows). The progression of JS is indicated by shaded arrow. Red box in the figure indicates the region of active convection during Dec, Jan and Feb while the violet box indicates the region of severe convection during Jan.

Table 5.3 CSE scale criteria for classifying CSE and their progression based on monthly averaged PD (hPa) and Jet Streak (m/s) from 1979-2017.

Category	Intensity of PD (hPa)	Strength of Jet Streak (m/s)	
Strong CSE	19-28	<=67	towards SCS
		68-77	towards SCS, Korea, Japan (bifurcation)
		<=78	towards eastern China, Korea, Japan
Extreme CSE	29-38	<=67	towards SCS
		68-77	towards SCS, Korea, Japan (bifurcation)
		<=78	towards eastern China, Korea, Japan

Table 5.4 lists the reported strong CSEs that resulted in severe socio-economic disasters during the years 2016, 2012, 2011, 2009, 2008, 2005, 2001, 1981 and 1968 in East Asia as tabulated in the Emergency Events Database (EM-DAT). Additional information on the occurrence of CSE in Table 5.4 have also been sought from web sources of Japan Meteorological Agency, Korean Meteorological Administration and HKO. It is noted that EM-DAT data contains only those surges over 1968 to 2016 that have resulted in large socio-economic disasters (e.g. losses in USD, countries affected etc.). Hence, there is likely additional cold surge events during the same period but these surges have not been reported by EM-DAT in that they have not caused large socio-economic losses and thus were minor surges. The categorization of the CSE as ‘Strong’ or ‘Extreme’ as listed in Table 5.4 is based on the CSE scale (Table 5.3). Extreme and Strong CSEs in Table 5.4 are based on intensities of PD. For these CSEs, the strength of Jet Streak as based on CSE scale (Table 5.3) provides an estimation of the progression of CSE track which was majorly towards the SCS during Dec 2009, Feb-2008, Dec 2001, Feb 1968; majorly

towards Japan during Jan 2011, Dec 2005, Jan 1981, Jan 1984; and bifurcation for Jan 2016, Jan 2008. The table also lists the computed monthly average PD between the SMH and AL systems. The ‘High’ and ‘Low’ values of PD, Jet Stream, SH & AL in the table indicates the value above and below the climatological mean respectively. For all the reported CSEs, the SMH is found to be higher than its climatological mean for all years except 2009 similarly the AL is found to be lower than its climatological mean for these reported years except Jan 2008. The higher values of Jet Streak indicate intense upper level divergence leading to atmospheric convection at the surface, triggering formation of LPS. Most importantly for all these events, PD is always found to be higher than its climatological mean indicating unusual atmospheric general circulation pattern as originating due to interaction of SMH and AL over East Asia, indicating a strong and persistent pressure system leading to CSEs. Notably the 2008 CSE, in addition to having a higher value of PD, also had the highest westward longitudinal shift in the AL of 22.5° from its climatological mean. This allowed an abundance of cold air to stream into mainland China, leading to the CSE to be unprecedented in duration and spatial extent. The above analysis suggests that the development of an intense PD dominates the progression of a continuous and intense band of winds from the Arctic Circle allowing it to reach far into the lower latitudes while the intensity of Jet Streak provides a guidance for the progression of CSE and thus the CSE scale can be considered as reliable criteria for predicting CSEs in East Asia. Thus, in summary, the PD information predicts the intensity of surges and the strength of JS determines the progression of the surges, thus providing a more scientifically sound method in predicting the onset of strong surges.

Table 5.4. Reported CSEs, countries affected, losses in USD available from EM-DAT, Monthly avg. PD, Pressure Gradient, intensification of SMH & AL system and shifting of AL system. High and Low indicate value above and below the climatological mean.

Reported CSE	Category of CSE based on CSE scale	Countries affected	Total USD ('000\$)	Strengthening of SMH (hPa)	Intensification of AL System (hPa)	Avg. PD (hPa)	Jet Streak (m/s)	Shifting of AL System towards West (Deg.)
Jan-2016	Extreme CSE	China, Japan, Korea, Taiwan, Mongolia, Hong Kong	NA	1037.381 (High)	1004.163 (Low)	33.22 (High)	69.66 (low)	No
Feb-2012	Strong CSE	China, Japan	20200	1031.69 (High)	1009.79 (Low)	21.9 (High)	74.62 (High)	No
Jan-2011	Extreme CSE	China, Korea, Japan	281000	1040.35 (High)	1009.68 (Low)	30.67 (High)	82.00 (High)	Yes, by 10 deg.
Dec-2009	Strong CSE	Mongolia, Hong Kong	62000	1029.53 (Low)	1010.15 (Low)	19.39 (High)	60.16 (Low)	No
Jan-2008	Strong CSE	China, Hong Kong	21100000	1036.85 (High)	1012.33 (High)	24.52 (High)	68.75 (Low)	Yes, by 22.5 deg.
Feb-2008	Strong CSE			1031.34 (High)	1010.79 (Low)	20.55 (High)	62.52 (Low)	No
Dec 2005	Extreme CSE	Japan	NA	1037.16 (High)	1005.77 (Low)	31.38 (High)	78.01 (High)	No
Dec-2001	Strong CSE	Taiwan	NA	1036.29 (High)	1009.72 (Low)	26.57 (High)	65.37 (High)	No
Jan-1981	Extreme CSE	China, Korea, Japan,	NA	1033.96 (High)	1001.51 (Low)	32.45 (High)	81.28 (High)	No
Feb-1968	Strong CSE	China, Hong Kong	NA	1036.93 (High)	1008.42 (Low)	28.52 (High)	65.65 (Low)	No
Jan 1984	Strong CSE	Japan	NA	1033.29 (High)	1011.53 (High)	21.76 (High)	77.78 (High)	No

Chapter 6

Conclusions

This study provides the first in-depth analysis on the relative roles of three major atmospheric systems, the SHM, AL and JS on the occurrence of CSEs during the NEM period in East Asia. We demonstrated that during the month of Dec, Jan and Feb as the widespread cold winds from the Arctic region reaches Mongolia through Siberia, it bifurcates towards lower latitudes impacting regions of South China, Vietnam, Taiwan, and towards eastern longitude impacting regions of eastern China, North-South Korea and Japan. Our investigation shows that the formation of intense and short lived LPSs between coast of Japan and western Pacific leads to this bifurcation. The analysis is performed using a combined approach of high resolution WRF-ARW model, Reanalysis data from ECMWF & NCEP, satellite data from ASCAT and station data from HKO, this being one of the very few such comprehensive studies.

Based on the detailed surge analysis and WRF modelling for South China (Chapter 4) and large-scale pattern of the surges over East Asia where we identified three new signatures related to the onset of CSEs comprising of SH, AL and JS (Chapter 5), we thus propose a new theoretical framework that explains the mechanism, the meteorological drivers and the process that leads to the development and formation of CSEs in East Asia. With the proposed framework we conclude a prolonged enhanced PD only develops due to the strengthening of SMH system, intensification of AL system under continued influence from JS results in strong CSEs in East Asia. In conclusion, we summarize the following main results from our study:

- i) The development of an intense PD dominated the progression of a continuous band of cold winds from Arctic Circle to Siberia reaching into the lower latitudes. The strong CSEs were accompanied by an enhanced and/or prolonged PD due to

strengthening of SMH and persistence intensification of AL. The former provided the persistent outbreak of cold air masses while the latter was crucial to lead the cold air masses as continuous cold advection.

- ii)** The intensity of PD is associated with the intensification of AL system, which is significantly influenced by the presence of JS. Moreover, the persistent pattern of intensification of AL occurs due to the strengthening of embedded Jet Streak within JS during Dec, Jan and Feb. Furthermore, strengthening of the Jet Streak leads to intense convection in the atmospheric column underneath the Jet Streak exhibiting rapid progression of CS from SMH.
- iii)** In addition to the strengthening of the Jet Streak, the pattern of JS is found to be another significant indicator for the onset of unusual long persisting extreme cold advection. For CSE Jan-Feb 2008 a very distinct pattern of zonal JS is observed for a continuous period of 14 days where the JS was observed to be elongated and stretched extending over a longitudinal length of 110°E . This indicates large upper air temperature difference through the troposphere resulting in a rapid progression of cold air from SMH at the upper level. Furthermore, the strengthened Jet Streak embedded within JS enhanced the intensification of the AL system resulting in a prolonged enhanced PD, leading to a record breaking cold advection in South China. Over this duration, HKO recorded the longest cold spells since 1968.
- iv)** The strength of Jet Streak and the pattern of JS determine the formation of intense short lived LPSs near the western coast of Japan and eastern pacific during Dec, Jan and Feb. Additional analysis of 38 years of data during these months further manifests the role of diverging air masses from Jet Streak in inhibiting the formation and development of LPSs. The formation of LPS leads to the bifurcation of the cold air mass from SMH into two major CS tracks that progress towards lower latitudes and eastern longitudes.

- v) CS track originating from SMH bifurcates into two major tracks under the combined effect of the pressure gradients that develops between SMH-SCS and SMH-LPS. The continuous strengthening of SMS-SCS gradient above its threshold value of 0.78 Pa/km over 18 hr duration and subjected to a SMH-LPS pressure gradient not exceeding 1.59 Pa/km leads to onset of CS over HK.

The occurrence of unusual CSE, as it severely impacts people and the economy is rare as the probability of occurrence of such CSEs is low. Therefore, the mechanism that leads to CSEs is not only of scientific interest but also of particular use for the decision makers who need reliable predictions of CSEs with extended duration. Over recent decades, although numerous scientific and technological improvements have been made to improve the forecasts, predicting extreme events has always been a challenge as the underlying physical mechanism still needs to be fully explored. Thus, it is important to investigate and characterize past events to improve our understanding of extreme weather. Towards this end, this PhD research has demonstrated a framework for identifying and characterizing the process that causes CSEs in East Asia.

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Appendix A: Additional Information on Chapter 3

To investigate the sensitivity of cumulus parameterizations scheme during the onset of CSs, we conducted WRF numerical experiments using four different parameterization and schemes and also by switching off the parametrization scheme. All these schemes captured the favorable atmospheric conditions with respect to surface pressure, temperature and wind as needed during the onset of cold surge (Figure A.1). However, though these schemes were able to capture the behavior of the atmospheric variables during the onset, a key issue was the triggering the convection during the onset of surges at 09 UTC on 15th Dec.

KF (Kain-Fristch), BMJ (Betts-Miller-Janjic), GD (Grell 3D Ensemble) and No_CU (No cumulus) triggered the cold surge MSLP values at 1017.50 hPa, 1017.53 hPa, 1017.42 hPa and 1017.30 hPa, respectively. These are lower values than that attained by GF (Grell-Freistas) scheme at 1017.62 hPa. Moreover, temperature started dipping suddenly after attaining 19 °C in GF scheme while the other schemes simulated comparatively a higher temperature ranging from 21-22 °C before triggering the step drop in temperature. Furthermore, all the schemes delayed in triggering the sudden change in wind speed from about 09 UTC except for the GF scheme which captured its early onset at 06 UTC. This employed GF scheme is the latest cumulus parameterization scheme updated in WRF V3.6 developed specifically for local convection and been tested for high a very high resolution (<10 km) valid for mesoscale models with horizontally varying grid resolutions (Grell et al., 2014). The other schemes could not reproduce triggering conditions behavior consistent with the observed data is probably due to lack of moisture in the surrounding. Since, the environmental profiles of the domain are already dry and cold, this may have limited the effectiveness of the schemes in simulating surges. This could also be due to the early release of latent heat and the resulting struggle to regain values similar to the other simulations. Another possibility is the lack of low-level convergence to support the continued growth of the initial and split updraft. However, a full explanation requires a further detailed investigation. Most importantly, these simulations demonstrated that caution must be used when employing a scheme for simulating surges. Two important conclusions are drawn from the simulations. Firstly, for small grid spacing (<10 km) is

not always appropriate to assume that model simulation may not need a cumulous parameterization scheme. Secondly, schemes other than GD scheme show a weaker representation of the surge onset and this probably due to strong dependence on available moisture. This GD scheme is used in the current thesis work.

Figure A.1. WRF, three hourly D2-plot at HKO station ($22^{\circ}18'07''N$; $114^{\circ}10'27''E$) from 15th Dec-16th Dec at 00 UTC for (a) for MSLP (hPa); (b) for 2m Temperature ($^{\circ}C$); and (c) for 10m Wind (m/s) for MSLP. Simulations were performed using KF, BMJ, GD, No_CU and GF schemes.

Appendix B: Additional information on Chapter 4

Table B.1. Monthly mean values and Normal (1981-2010) for Jan 2016 for surface pressure, min and mean temperature, mean wind speed and total rainfall at HK.

Jan 2016	Daily Mean Pressure (hPa)	Daily Min Temp (°C)	Mean Temp (°C)	Mean Wind Speed (km/hr)	Total Rainfall (mm)
Mean/Total	1020.4	14.4	16	29.4	266.9
Normal	1020.3	14.5	16.3	25.3	24.7

Data Source: HKO

Table B.2 Monthly mean values and Normal (1971-2000) for Jan 2008 for surface pressure, min and mean temperature, mean wind speed and total rainfall at HK.

Jan 2008	Daily Mean Pressure (hPa)	Daily Min Temp (°C)	Mean Temp (°C)	Mean Wind Speed (km/hr)	Total Rainfall (mm)
Mean/Total	1019.1	14	15.9	22.7	33.3
Normal	1020.1	14.1	16.1	25.4	24.9

Data Source: HKO

Table B.3 Monthly mean values and Normal (1971-2000) for Feb 2008 for surface pressure, min and mean temperature, mean wind speed and total rainfall at HK.

Feb 2008	Daily Mean Pressure (hPa)	Daily Min Temp (°C)	Mean Temp (°C)	Mean Wind Speed (km/hr)	Total Rainfall (mm)
Mean/Total	1020.7	11.3	13.3	25.6	27.5
Normal	1018.6	14.4	16.3	25.1	52.3

Data Source: HKO

Table B.4 Monthly mean values and Normal (1971-2000) for Dec 2009 for surface pressure, min and mean temperature, mean wind speed and total rainfall at HK.

Dec 2009	Daily Mean Pressure (hPa)	Daily Min Temp (°C)	Mean Temp (°C)	Mean Wind Speed (km/hr)	Total Rainfall (mm)
Mean/Total	1020.7	11.3	13.3	25.6	27.5
Normal	1020.5	15.6	17.7	26.3	38.34

Data Source: HKO

Figure B.1: HKO daily forecast for Temperature for CSE 2016. In the figure European model refers to ECMWF, Japan model refers to JMA, US model refers to GFS and actual observation refers to observed data recorded at HKO.

Source: <https://www.hko.gov.hk/press/SP/pre20160125.htm>

Figure B.2: HKO daily MSLP (in black) and WRF-D1 daily MSLP (in red) from 24th Jan to 16th Feb 2008 at 00 UTC.

Appendix C: Additional information on WRF Model diagnostics

To investigate the impact of PBL during the onset of CSs simulations were completed using 9 x 9 km horizontal grid spacing over South China with domain center at HKO. The use of PBL schemes enables parameterization of the unresolved turbulent vertical fluxes of heat, moisture and momentum within the PBL and throughout the atmosphere. Here a closure scheme is essential to derive turbulent fluxes from mean quantities (Holt and Raman 1988). The height of PBL provides important information for local convection and has been used as a key parameter (e.g. Shin et al., 2007). We have used the YSU PBL scheme (Hong et al. 2006) which is a first order nonlocal scheme. This scheme has been modified in WRF version 3 by increasing the critical bulk Richardson number from zero to 0.25 over land, thereby enhancing mixing in the stable boundary layer (Hong and Kim 2008).

For cold surge 2008, the earlier period from 24th Jan to 4th Feb had the cold spell mostly driven by strong winds flowing through the Straits of Taiwan towards HK. During this period the daily PBL height continually remained below 200 m starting from 26th Jan to 3rd Feb (Figure C.1a). The period from 5th Feb to 16th Feb had cold air arriving from Mongolia towards mainland China as a CS reaching HK via north easterly winds. During this period many fluctuations in PBL height was observed. However, the PBL height suddenly increased from 200 m on 6th Feb to 600 m on 9th Feb and suddenly decreased from 650 m on 13th Feb to 100 m on 14th Feb. It peaked above 700 m on 16th Feb.

For Cold Surge 2009, the surge duration was between 15th-16th Dec 2009. During this period with the arrival of the surge on 15th at 9 UTC there was a sudden increase in PBL from 150 m to 450 m (Figure C.1b). With the passage of the surge, the PBL started to decrease continually from 450 m from 11 UTC to 200 m till 18 UTC. For Cold Surge

2016, with the arrival of the surge, a sudden increase in PBL is noticed from 475 m at 23 UTC on 23rd Jan to 700 m at 5 UTC on 24th Jan (Figure C.1c).

The current understanding of the temporal change of PBLH in relevance to the onset of CS is very limited. Therefore, it is probable that these sudden increase in the PBLH could be interpreted as enhancement of the convective mixing within the boundary layer. This enhanced mixing could result due to advection of strong winds associated with onset of CS. Such convective mixing would probably lead to expansion of PBL. While with the continued arrival of abundance of cold air, these convective mixing later starts to weaken and PBL starts to contract. However, this particular behavior of PBLH requires a further detailed investing in understanding its possible role in relevance to cold surge onset.

Figure C.1: (a) Daily PBLH during Cold Surge 2008, (b) Hourly PBLH during cold surge 2009 and (c) Hourly PBLH during Cold surge 2016 at HKO (Lat:22° 18' 07", Lon: 114° 10' 27"). Horizontal axis in (a) indicates days from over 25th Jan-16th Feb 2008 at 00 UTC. Horizontal axis in (b) indicates hours from 15th Dec 07 UTC to 18 UTC and (c) indicates hours from 23rd Jan 23 UTC 24th Jan 10 UTC.

Appendix D: Supplemental Figures for Chapter 5

ECMWF MSLP (hPa) at 00 UTC on 12th Jan 2016

ECMWF MSLP (hPa) at 00 UTC on 13th Jan 2016

ECMWF MSLP (hPa) at 00 UTC on 15th Jan 2016

ECMWF MSLP (hPa) at 00 UTC on 19th Jan 2016

Figure D.1. ECMWF, ERA Interim plot for MSLP (a) on 1st Jan 2016 at 00 UTC; (b) on 5th Jan 2016 at 00 UTC; (c) on 7th Jan 2016 at 00 UTC; (d) on 11th Jan 2016 at 00 UTC; (e) on 12th Jan 2016 at 00 UTC; (f) on 13th Jan 2016 at 00 UTC; (g) on 15th Jan 2016 at 00 UTC; (h) on 19th Jan 2016 at 00 UTC; (i) on 20th Jan 2016 at 00 UTC; (j) on 25th Jan 2016 at 00 UTC; and (k) on 31st Jan 2016 at 00 UTC shows formation of frequent LPS during Jan-Feb 2008 between 140°E - 180°E and 30°N - 55°N.

ECMWF MSLP (hPa) at 00 UTC on 4th Dec 2009 ECMWF MSLP (hPa) at 00 UTC on 13th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 14th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 16th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 17th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 18th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 19th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 20th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 21st Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 22nd Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 27th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 28th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 29th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 30th Dec 2009

ECMWF MSLP (hPa) at 00 UTC on 31st Dec 2009

Figure D.2. ECMWF, ERA Interim plot for MSLP (a) on 2nd Dec 2009 at 00 UTC; (b) on 4th Dec 2009 at 00 UTC; (c) on 7th Dec 2009 at 00 UTC; (d) on 13th Dec 2009 at 00 UTC; (e) on 14th Dec 2009 at 00 UTC; (f) on 16th Dec 2009 at 00 UTC; (g) on 17th Dec 2009 at 00 UTC; (h) on 18th Dec 2009 at 00 UTC; (i) on 19th Dec 2009 at 00 UTC; (j) on 20th Dec 2009 at 00 UTC; (k) on 21st Dec 2009 at 00 UTC; (l) on 22nd Dec 2009 at 00 UTC; (m) on 27th Dec 2009 at 00 UTC; (n) on 28th Dec 2009 at 00 UTC; (o) 29th Dec 2009 at 00 UTC; (p) on 30th Dec 2009 at 00 UTC; and (q) on 31st Dec 2009 at 00 UTC shows formation of frequent LPS during Jan-Feb 2008 between 140 0E - 180 0E and 30 0N - 55 0N.

ECMWF MSLP (hPa) at 00 UTC on 7th Jan 2008 ECMWF MSLP (hPa) at 00 UTC on 10th Jan 2008

ECMWF MSLP (hPa) at 00 UTC on 3rd Feb 2008

ECMWF MSLP (hPa) at 00 UTC on 4th Feb 2008

ECMWF MSLP (hPa) at 00 UTC on 7th Feb 2008

ECMWF MSLP (hPa) at 00 UTC on 8th Feb 2008

Figure D.3. ECMWF, ERA Interim plot for MSLP (a) on 6th Jan 2008 at 00 UTC; (b) on 7th Jan 2008 at 00 UTC; (c) on 9th Jan 2008 at 00 UTC; (d) on 10th Jan 2008 at 00 UTC; (e) on 3rd Feb 2008 at 00 UTC; (f) on 4th Feb 2008 at 00 UTC; (g) on 7th Feb 2008 at 00 UTC; (h) on 8th Feb 2008 at 00 UTC; (i) on 10th Feb 2008 at 00 UTC; (j) on 11th Feb 2008 at 00 UTC; and (k) on 13th Feb 2008 at 00 UTC shows formation of frequent LPS during Jan-Feb 2008 between 140°E - 180°E and 30°N - 55°N.

Appendix E: Additional information on Chapter 5

Table E.1. Daily mean for Jan 2016 and Monthly mean values for Jan (1979-2016) for SMH, AL and JS. The shaded region shows the period of the cold surge onset.

Jan 2016	PD (hPa)	JS-Daily Mean Wind Speed (m/s)
1	23.049	56.13
2	28.316	55.23
3	33.674	57.45
4	33.071	60.49
5	30.209	63.23
6	38.927	64.45
7	33.152	62.11
8	26.77664	59.28
9	27.133	58.67
10	34.526	58.00
11	35.381	56.92
12	32.937	57.42
13	29.073	59.12
14	23.641	57.17
15	21.242	55.79
16	30.722	53.41
17	30.263	53.87

18	26.413	54.41
19	26.870	51.97
20	31.125	55.12
21	39.756	56.48
22	49.268	57.05
23	49.437	59.23
24	44.946	62.71
25	39.748	61.38
26	35.011	63.23
27	31.613	65.97
28	37.442	67.03
29	38.843	65.30
30	34.826	64.34
31	32.350	65.84
Daily Mean of 31 days	33.1	59.32
Climatology (1979-2016)	21.1	70.6

Data Source: ECMWF

Table E.2. Monthly mean values and Normal (1979-2016) for Jan 2008 for SMH, AL and JS. The shaded region shows the period of the cold surge onset.

Jan 2008	PD (hPa)	JS-Daily Mean Wind Speed (m/s)
1	19.507	49.55
2	22.618	49.77
3	24.849	51.03
4	25.195	51.48
5	21.749	53.31
6	19.668	55.82
7	18.983	55.36
8	18.317	57.49
9	15.597	59.05
10	18.848	57.95
11	26.951	55.55
12	34.077	55.035
13	33.648	54.53
14	32.452	58.24
15	31.652	57.60
16	27.650	58.16
17	21.515	59.74
18	24.273	55.75
19	28.664	54.06

20	31.036	54.19
21	25.594	53.60
22	22.389	53.40
23	24.413	52.53
24	21.770	50.95
25	24.173	50.30
26	27.176	50.52
27	25.874	50.75
28	24.530	53.55
29	22.505	53.63
30	22.170	51.94
31	22.302	52.52
Daily Mean of 31 days	24.6	54.1
Climatology (1979-2016)	21.1	70.6

Data Source: ECMWF

Table E.3. Monthly mean values and Normal (1979-2016) for Feb 2008 for SMH, AL and JS. The shaded region shows the period of the cold surge onset.

Feb 2008	PD (hPa)	JS-Daily Mean Wind Speed (m/s)
1	19.136	53.560
2	17.394	54.554
3	15.490	53.329
4	19.728	53.149
5	24.354	52.808
6	24.629	53.075
7	23.490	54.214
8	20.121	52.263
9	18.942	53.341
10	29.031	54.510
11	31.043	55.82
12	25.504	55.52
13	22.682	55.98
14	20.786	56.52
15	24.194	57.88
16	26.168	56.98
17	28.132	54.83
18	33.093	51.89
19	27.736	49.05

20	23.330	51.70
21	16.276	53.42
22	15.840	56.29
23	19.912	55.46
24	18.675	51.80
25	13.994	51.67
26	11.912	52.68
27	12.919	52.12
28	4.847	53.64
29	6.651	55.53
Daily Mean of 29 days	20.6	53.9
Climatology (1979-2016)	16.6	68.3

Data Source: ECMWF

Table E.4. Monthly mean values and Normal (1979-2016) for Dec 2009 for SMH, AL and JS. The shaded region shows the period of the cold surge onset.

Dec 2009	PD (hPa)	JS-Daily Mean Wind Speed (m/s)
1	20.420	49.93
2	21.346	52.58
3	20.0966	53.95
4	12.390	52.36
5	2.2562	52.26
6	7.360	50.21
7	11.023	46.97
8	14.272	46.27
9	12.265	45.46
10	14.131	47.27
11	19.144	48.10
12	20.801	46.80
13	20.201	46.33
14	26.036	47.25
15	30.040	50.45
16	31.919	53.42
17	30.300	56.69
18	34.370	58.65
19	28.404	58.71

20	21.278	58.48
21	15.486	58.57
22	17.284	56.67
23	21.627	54.17
24	32.639	52.03
25	24.525	51.87
26	24.626	50.79
27	22.328	51.35
28	22.460	52.77
29	29.710	51.93
30	24.350	52.20
31	14.548	56.69
Daily Mean of 29 days	20.9	51.9
Climatology (1979-2016)	18.9	62.3

Data Source: ECMWF